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СЕРИЯ ГЕОЛОГИЧЕСКАЯ

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NEW CONCEPT OF TERTIARY OROGENESIS OF THE PYRENEES[1,2]

by

J. P. MANGIN

It is very difficult to obtain a complete pic-
e of Paleogene paleogeography for the Py-
nees province of France and Spain. Nowhere
s this topic been considered as a whole; the
st recent works, collating the results of
est observations, make a clean-cut dis-
ction between the two slopes of the Pyrenees,
accordance with the E. Haug concept that
he Paleogene series on the Spanish slope of
Pyrenees... is substantially different from
t on the French slope. It can be inferred
t the la ter was deposited in a geosyncline
te independent of that of the Front Pyrenees,
d evidently connected with the western Alpine
osyncline, by way of regions now submerged
the Mediterranean" ([14], p. 1468).

Despite the objections to such a thesis from
me students of both slopes (L. Carez, M.
lloni, G. Mengaud, R. Ciry), it appears that
ug's differentiation of the slopes has become
anon: it has been adopted in the construction
the latest paleogeographic maps.

Indeed, the present day crest of the Pyrenean
al trend appears to many authors to be a very
ient barrier. Because of this erroneous pre-
se, the study of the two parts of the Pyrenean
in as a discrete unit has ceased; judging
m the literature, one may think that the
two Paleogene geosynclines of E. Haug are indeed present.

I deem it necessary to summarize the stratigraphic data gathered separately for each slope of the Pyrenean range and use it as far as possible for apleogeographic conclusions valid for either slope; then to build on this data certain hypotheses on the Tertiary origin of the Pyrenees.

What are the data at our disposal?

On the French side, Paleogene stratigraphy of the Pyrenees has been known virtually since the beginning of this century. The study proceeded simultaneously in the west part of the Aquitanian plain and in east Aquitania (Garonne Supérieure Ariège). The correlation of various stratigraphic units was almost completed by E. Hebert and Meunier-Chalmar. The role of foraminifera and especially of nummulites was determined by H. Douvillé who set forth in several very precise papers (1905-1906) the principal features of Paleogene stratigraphy of west Aquitania and demonstrated its relationship to other basins of Europe. East Aquitania was the subject of a livelier discussion, with the age of its continental deposits being more controversial than other sequences.

At the same time, in 1911, on the basis of data from the north slope alone, E. Haug regarded the Pyrenees as a structure which had originated on the whole in "post-Lutetian" or "Pyrenean" movements which lasted from the Lutetian to the close of the Oligocene.

From their very beginning, these movements produced uplifts in the east, which, upon erosion, gave rise to the famous "Palassou puddings (conglomerate)". Subsequently these uplifts spread westward and led to the formation of the Pyrenees proper. As a result of this process, the abovementioned conglomerates themselves were involved in the uplift.

However, after a new study of nummulites, H. Douvillé modified the existing stratigraphic scale (1917, 1919) on the basis of controversial considerations on the evolution of the nummulite

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[1]Novye predstavleniya o tretichnom orogeneze
reneyev.

Hypotheses Nouvelles sur l'orogénèse Tertiare
Pyrenees, par Jean Philippe Mangin. This paper
been received under an agreement with Societe
logique de France for reciprocal poblication of
ks by French and Soviet geologists in Izvestiya AS
S.S.R., Geologic Series and Bulletin de Societe
ologique de France. (Editorial Board).
This paper develops concepts partially set forth
the author in previous publications by the Societe
ologique de France. It uses material from a
per read before the Third Congress on the study
the Pyrenees (Gerona, 1958), as well as new
a obtained from recent observations, which
roborate the author's view. (Author's note).

shell. Thus he placed the Gannes marl (south of Pau, Figure 1), formerly regarded as Lutetian, in the Lower Eocene because the similar Beau d'Arro series contained Nummulites lucasi believed to be less advanced in evolution than N. Laevigatus, typical of the Lower Lutetian. It was not taken into consideration that the Gannes marls contain assilinas and nummulites previously known from the Lutetian. So, in the future, these fossils will be regarded as typical even of the Lower Eocene! Since then, the Ariége marl with turritellas, also regarded as Lutetian, became Yprèsian for H. Douvillé, while the overlying Palassou conglomerate he assigns to the Lower Lutetian. As a result of all that, the Pyrenean range is "aging" in works of those authors who regard these clastic formations as a formal criterion for the beginning of Pyrenean folding.

As the age of the Pyrenean conglomerates changes depending on the school of thought of its students, so does the age of the overlying continental formations with their terrestrial and fresh-water faunas. These faunas came to be regarded as typical of the beds which contain them, after they had been described. This undoubtedly led to a vicious circle of concepts first noted in 1816 (by Pouchen); fossils were determined from their enclosing rock which were determined from the fossils!

So it came about that the first horizontal beds, unconformably overlying the Pallasou conglomerates, are often dated as Oligocene. This view ignores the earlier studies which noted a certain reworking of "index" detrital beds; and some publications, even recent ones, assign an Eocene age to the Pyrenean range, a controversial conclusion even for the north slope.

For the south slope, these considerations are invalid. The assigning by M. Dalloni of thick, vertically standing conglomerate beds to the Ludian Eocene stage forces that author to place the terminal sharp movements "at the boundary between the two periods" (Eocene and Oligocene). On the other hand, the German school, headed by H. Stille, notes that the bulk of the clastic series is Oligocene (and even Miocene). This was subsequently confirmed by Spanish geologists who found a Bartonian fauna, e. g., in Biarritz, in blue marl of considerable thickness, underlying the familiar Monterra conglomerate. As early as 1920 (L. Mengaud) the existence of marine Oligocene beds was recognized on the Cantabrian coast, correlative with Oligocene deposits of the Basque-French coast. The contradictory nature of these data renders quite impossible any satisfactory simple orogenic picture for the Pyrenees as a whole.

How is it possible for a mountain structure formed in Lutetian time on the French side to provide clastic material for an accumulation of conglomerate on the south side, beginning as recently as the Oligocene? This problem presents difficulties because of the present separate study of the two slopes of the Pyrenees; it appears that data presented by different schools of thought appear to be contradictory.

It should be noted that a comprehensive survey of the literature to which I was introduced by Prof. R. Ciry in 1952 has led me, too, to contradictory conclusions. According to most authors, the north Pyrenean slope was formed in Lutetian time or later, while an Oligocene and even later age is assigned to the south slope by Spanish and foreign geologists. What, then, is the truth?

The very first thing to do was to evaluate the composite stratigraphic section, compiled mostly from the H. Douvillé observations, in order to determine its applicability to the entire Pyrenean province. As a result, it was determined that stratigraphic units were of the same age on both slopes. A vertification of observations in Aquitania could be facilitaed by modern techniques of geologic study.

I was presented with a chance for such a verification during my study of the Paleogene in the southeast[3] sector of the Pyrenees [18].

This area, little known stratigraphically, contains, indeed, complete sections of deposits missing in Aquitania. Their good exposures and simple tectonics afford a chance for a detailed study of stratigraphy and facies changes and for generalizing the data so obtained for the Pyrenees as a whole.

As a basis for our study of geologic history in the area of Paleogene development, which is the subject of this paper (an area of 5200 km^2 between the Santander and Jaca meridians, Figure 1), we took the sequence of evolution of foraminifera, particularly the pelagic species, and the course of sedimentary processes. Thus, our work is based on micropaleontologic and lithologic studies; in combination, they have corroborated our stratigraphic differentiation. It appears that the simplest way to study the history of development of a structure like the Pyrenees is to determine its most important stages; to associate with them secondary events of local importance affecting the sedimentary processes within a limited area; and thus arrive at a specific "biotype". This approach is preferable to the description of a series of "formations" often presenting a veritable mosaic, especially in tectonically active regions where the sequence of "faunal zones" is not always directly related to the geologic history.

[3]Translator's Note: It appears from the context that "Southwest" sector is meant.

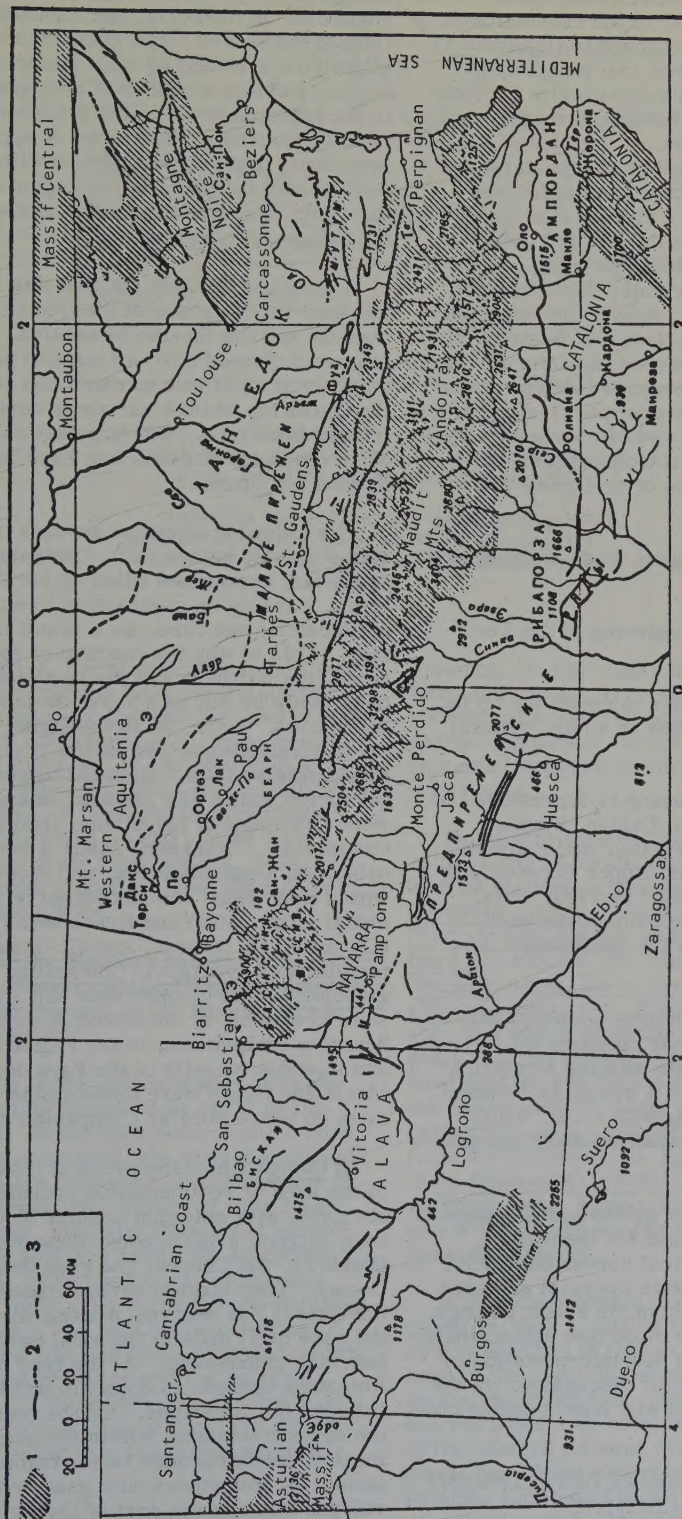

FIGURE 1. Principal points in the Pyrenean province. Structural plan of the north slope after M. Casteras; of the south slope, after L. Mengaud, Korenberg, R. Ciry and Ras (Asturia and Biscaya), and Rios and Almela (Catalonia).

1 - Paleozoic province; 2 - fault contacts; 3 - anticlinal axes.

After that, we attempted to relate these evants, as far as possible, to the conventional subdivision into stages, in order to illustrate the history as a whole. For that purpose, it became expedient to evaluate critically the identification of stratotypes, Paleogene stages, and typical Aquitanian sections, in the light of sedimentation and the progressive change in microfauna. On the whole, such a correlation was found feasible because of a certain similaritity, however tenuous, between conventional stages and various units observed in our area. However, in order not to undermine the historical value of the term "stage", despite the opportunity of describing here, in Navarre, some of the formerly indefinite stratotypes, I prefer for the time being to use general names (such as transitional unit or Upper Eocene proper) instead of some little-known names and new terms not approved by the International Commission, in order to avoid introduction of erroneous nomenclature.

The study of the Paleogene leads to three basic conclusions:

1) The Danian is the beginning of the Tertiary [16, 18];

2) There are no reasons to retain the term, Paleocene, except in studies of a most general nature [17, 21];

3) The Pyrenean Paleogene is represented by the following subdivisions: Danian, Montian, Landenian, Cuisian (Ypresian)), and Lutetian stages, a transitional unit, Upper Eocene, and Oligocene, the latter divided into four units, commonly with marine deposits at the base, possibly correlative with the Sanoissian-Stampian.

The first three of the stages named above, under continental conditions, became the Garumnien facies which begins with the Maestrichtian but does not belong to it alone, as believed by many authors beginning with L. M. Vidal. This is what made many students doubt the presence of the Lower Eocene.

I have considered these subdivisions before [20, 22] and shall not pause for them again. They constitute an historical canvas on which I have organized the extensive material gathered by my predecessors to whom the credit belongs. New data will be added to those now extant; their correlation may lead to a new interpretation or at least provide material for discussion. In the meantime, my hypotheses will stand pending confirmation or rejection.

HISTORY OF THE PYRENEAN DEVELOPMENT IN THE PALEOGENE

Toward the close of Maestrichtian time, the end of the Cretaceous period, the geography of the Pyrenean province was as represented in Figure 2. In the north, the Aquitanian platform fringed the Massif Central of France, possibly slightly rejuvenated by weal Laramide movements, and was a source of terrigenous material in the Maestrichtian. This platform extended from Charantes to Minervoix; it appears to have been connected in the southeast with the vast Ebro land located in Spain, in the region of the present-day Ebro, and connected the Asturian massif with the present east Pyrenees.

A large embayment cut into this low-lying land, elevated only in zones of the Central and Asturian massifs and the Iberian Mountains. The embayment was divided into two unequal parts: a wide continental shelf was an extension of the Aquitanian Platform, while a narrower neritic zone fringed Ebro land. The greatest subsidence occurred in the present-day Pays des Basques of France and Spain.

The following situation appears to have prevailed shortly before that: Campanian limestone with Hyppurites filled up the Pyrenean Strait, causing its contraction to an embayment, in Maestrichtian time, as a result of a regression. This is why the junction of the two continents, the northern and the southern, was effected only by way of a very low land with mobile boundaries. Here, as in the higher land area, the Garumnien-type facies was deposited during Maestrichtian time. Reduced to simplest terms, the facies within this nascent "eastern isthmus" were first lagunal lacustrine, then strictly continental. "Nankeen" sandy limestone with orbitoids, omphalocycles, and siderite, locally with a marl lens, was deposited on the continental slope. Marl and flysch with Glogotruncana (Gl. arca, Gl. contusa, and Gl. stuarti), Pseudotextularia (P. varians), Gümbellina (Heterohelix globulosa), and globigerinas (Gl. gr. massinae) were prominent at the mouth of the Biscay-Bearnian embayment, in the region of ancient and unstable massifs of the Pays des Basques. Marls and flysch were deposited in this embayment on top of similar Campanian deposits.

In Danian time, the influx of clastic material to the area of sedimentation probably ceased abruptly, with only thin nodular limestone, locally slightly argillaceous, deposited everywhere (Navarre, the Landes, the Pays des Basques Coast). They contain several index species of the typical Fax section, present also in some of the rare marl beds in the axial part of the Biscay-Bearnian embayment. At the land margin, these limestone changes to dolomite and locally to lithographic limestone. On the continent and on the eastern isthmus, within the area of development of the Garumnien facies comples, mostly lacustrine limestones with gastropods were developed (the eastern part of the Aquitanian plain, Catalonia, Navarre, and the Aragonian Sierra).

Parallel to this abrupt termination of the influx

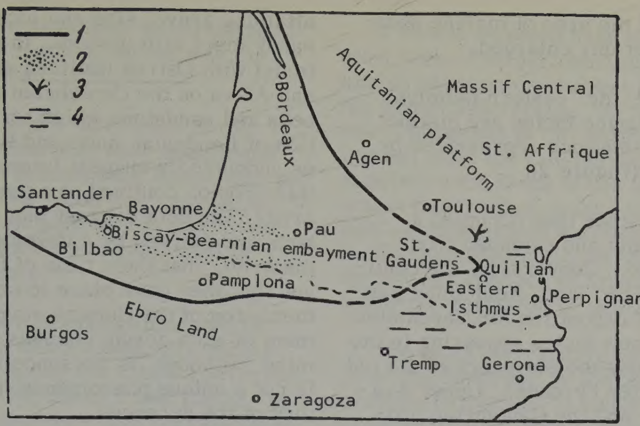

FIGURE 2. Paleogrogeaphic map at the onset of the Paleogene, from the end of the Maestrichtian to the Cuisian; after J.P. Mangin, 1959.

1 - Inferred shore line; 2 - pelagic provinces, generally shallow, with marl and flysch; 3 - lacustrine facies; 4 - Thanetian marine stage of the eastern isthmus.

f clastic material, the marine fauna underwent radical change; judging from foraminifera, the auna became impoverished and included only a arrow range of Globigerinas (G. pseudobullo-des, G. triloculinoides, G. danbjergensis, G. Globorotalia? compressa) and some miliolids and tubercularias, without the former abundance of warmth-loving Maestrichtian forms. Obviously, the climate grew cooler. It is possible that some secondary land species persisted in isolated places.

In any event, the Tertiary had arrived. Unlike the abrupt Maestrichtian-Danian transition, he Danian stage changed to the Montian gradually and imperceptibly.

The filling up of the sedimentation province probably was fairly advanced as early as the Danian because, as we have seen, sedimentary facies here are comparatively little diversified. During the Montian, however, the deposition of uniform algal limestones occurred over large but shallow areas, virtually throughout the entire submerged province within the Maestrichtian boundaries. Algal and coral limestones overlie Danian marl even in the Biscay-Bearnian embayment zone. The lush development of lithothamnias and microdiaceae extended as far as the continental province and the eastern isthmus where the Vitrollien deposits (part of the Garumnien facies) consist partly of the remains of algae living in a lacustrine environment.

That probably was brought about by climatic changes and is in full accordance with the appearance of similar facies in the boreal province. It was under these conditions that Montian deposits were widely distributed over the submerged area (Figure 2); the widely known Ariège deposits in the Lower Pyrenees and the Santander deposits in the Perdido Mountain occur here.

Three stratigraphic units are identified here on the basis of foraminifera: 1) a basal horizon with an abundance of algae and miliolids (Planorbulina antiqua) and some large foraminifera Discocyclina seunesi and Operculina pseudoheberti; 2) another unit with algae and Rotalia trochidiformis, Eponides gr. lotus, Rotorbinella colliculus, etc; 3) a unit where algae are rare and where miliolids predominate with some species of Miscellanea and roalids. Globigerinas are well represented throughout the entire section; along with Globigerina triloculinoides and G. Pseudobulloids, inherited from the Danian stage, there are similar groups of G. imitata, G. perclara, G. Elongata, G. varianta, etc., known from some very rare marly intercalations (as in the Tercy Ridge, Spanish Navarre).

It is quite probable that Montian deposition coincided with a marine transgression, evidenced by a pebble bed of Danian rocks occurring at the base of the algal limestone and resting directly on the Maestrichtian. Furthermore, the Montian marine basin may have included the western part of the eastern isthmus where lagunal or marine facies have been observed in many places (east Aragon, Catalonia) in the area of development of redbeds which represent the Montian in the Garumnien sequence. It can be assumed that, toward the close of the Montian, the sea spread

over the continent and the area of marine sedimentation was considerably enlarged.

In the Landenian,[4] the "eastern isthmus" was submerged and marine facies are distinct above the continental sequence represented by the Garumnien facies (Figure 2).

The miliolid limestone then occupied a sizable area of Catalonia and extended from Perpignan to Moutoumé. Conversely, the continental areas proper (Ebro land, Aquitanian platform) because sites of deposition for Garumnien type sediments, with new facies appearing in the submerged parts. Along the periphery of the old land (the Landes, Petites Pyrenees, Upper Aragon, Navarre, Alava, and the Cantabrian province), Landenian deposits are represented by miliolid limestone with the first alveolines (Alveolina ovulum, A. primaeva) and often with Fallotella alavensis, Miscellanea, and Asterodiscus taramellini. These limestones differ little from the Upper Montian to which they are very similar in facies. Miliolids in them are more numerous, while the appearance of the first alveolinas marks a clean-cut although arbitrary boundary. Conversely, going toward the Landenian province, farther away from the shore, deposits are represented by a facies of marl and sandy flysch with the first truncorotalids, especially Truncorotalia angualate, T. aequa, Globorotalia elongata, Tr. acuta, Tr. velascoensis, and Globigerina eocenica.

These deposits rest on Montian limestone and occupy a large area of the expanded ancient Biscay-Bearnian embayment. The appearance of terrigenous elements after a marine epoch probably suggest a slight recurrence of epeirogenic movements which set in motion the detritus which originated in Danian and Montian time and shifted them to the bottom, now again submerging them after a pause at the beginning of the Paleogene. The deeper parts were deposited approximately along the present axial zone, west of the Ori peak, except for some bottom highs where miliolid limestones were deposited (south of Saint Jean Pied de Port).

The structure of the present day Pays des Basques is, then, somewhat mosaic-like, with some Eocene segments depressed and others exposed. This is quite natural because such a structure is typical of the Pyrenees as a whole. Thus, the Landenian facies are related to local features of the bottom.

The close of the Landenian was represented throughout the Pyrenean province by the short-lived appearance of coarse clastic material, attaining gravel size and locally changing to sandy marl with oysters, in the littoral zone (marl with Ostrea uncifera at Aude, Catalonian and Alava on the Cantabrian coast). The gravel beds and sandstone which terminated the deposition of Landenian marl and flysch in the Pyrenees undoubtedly suggest intensive erosion. Was this erosion confined to a possible continental area? The data suggest that terminal Landenian beds are marked by a development of clastic material, that their time of formation was brief, and that they gave place to transgressive sedimentation of the Cuisian stage. The environment of such abrupt changes is difficult to determine, although its presence is quite definite and is not a unique phenomenon in the geologic history of the Pyrenees.

The great Cuisian transgression.[5] In most places, the Cuisian sedimentation proceeded gradually. Resting on Landonian sandstones are thin transitional limestones, still carrying some of the clastic material; then the excellent alveolina limestone, widely developed throughout the Pyrenees appears suddenly. A new Pyrenean strait came in to existance, similar to that of the Campanian. The shores receded and without any disturbance the Cuisian sea flooded the low stretches of the coast: south of Mt. Noire, the Gerona province, the Ebro land periphery, and the remains of the eastern isthmus. Myriads of alveolina and miliolina shells (A. oblongata, A. subpyrenaica) were deposited everywhere in shallow waters, to form this familiar limestone, but were locally replaced by coralline limestone.

The transgression damped the influx of terrigenous material, and the limestone became virtually free of impurities. It was deposited over the entire newly submerged area; it also caps continental deposits of the Garumnien facies. The distribution area of these characteristic limestones is very broad, with littoral facies very rarely present. Finely clastic limestone with Nummulites globulus, N. quettardi, and N. atacicus, some star-like discocyclinas, Asterodiscus taramelli, and the first Discocyclina archiaci, occur here and there, suggesting that their deposition took place in a bottom environment too mobile for the deposition of alveolina limestone.

In those regions where alveolina limestones are fully represented, provided they do not carry nummulites as well, strictly Lutetian assilinas and nummulites are never present. Despite the fact that this facies is very favorable for them, they appear only in much higher horizons. The thick marls (Gannes, Cuise, etc.) which carry these foraminifera and which owe their origin to the influx of terrigenous material unknown in

[4]The name Landenian is sometimes given to beds corresponding to the Thanetian and Sparnacian stages, i.e., to the top of the Paleocene. Russian Editorial Board.

[5]Cuisian stage, a synonym for the Ypresian, consists of an essentially sandy Ypresian facies (Lower Eocene). Russian Editorial Board.

uisian beds should be assigned, in my opinion, the Lutetian. This opinion, by the way, is hared by other authors.

True Cuisian marls are quite rare in the yrenean province where they have been found nly as intercalations in alveolina limestone, at linervoix. It should be noted that the distri-ution of facies is often difficult to trace in de-il, because much has been camouflaged by ost-Cuisian movements.

Cuisian deposits are always culminated in ne gravel (like those in the Landenian stage) bserved everywhere along the Cantabrian coast, t Mt. Noire, in the Pyrenean Front Range, in avarre, etc. This evidence of a regression is ollowed by Lutetian marl and limestone.[6]

The Lower Eocene presents a single unit; for his reason, the conception of the Paleocene and ot the Lower Eocene being represented by the uisian alone, is obsolete. Nothing separates he Landenian and Cuisian faunas. The sedi-entary and paleontologic ties have their roots the Montian which succeeds the Danian.

The regression which terminated the Early ocene was preceded by movements of momen-ous import in the geologic history of the Py-enees; they were the beginning of a slow de-elopment of the mountain range. They were he harbingers of the Tertiary restoration of an le Hercinian structure. Broad gentle folds, ending parallel to the shores of the strait ere formed then, the future Pyrenean trend rom west-northwest to east-southeast). At the ame time, in the eastern part of this province, his trend followed that of ancient massifs, the It. Noire and Catalonian.

The gentle flexing of the beds, caused either y some rapprochment of the confining massifs r by a subsidence of the bottom (or by both of hese and allied causes), led to two considerable hanges; the continental province was enlarged y the addition of two new regions, the Mt. Noire nd the Asturian marginal zone, and the present ea bottom, fairly monotonous in Cuisian time, as dismembered, which is in accordance with he concept of an abrupt subsidence of individual lements of the axial zone.

Such were the changes which took place dur-ng late-Cuisian to early Lutetian movements. 'or convenience, I shall call them the early utetian phase. In many places, this phase is robably equivalent to the "Pyrenean paroxysm" f some authors. It follows from what has been aid that this "paroxysm", despite its obviously mportant role, did not uplift the mountain range tself.

[6]Middle Eocene. Russian Editorial Board.

It should be noted here that these events could have been of rather short duration because there are no perceptible changes in the development of the faunal assemblage, so that Lutetian faunas which appeared in the sedimentary province deformed after the Cuisian deposition are very similar to the fauna of the last Lower Eocene stage; the Lutetian faunae may be their direct and uncontaminated successor. To be sure, the facies themselves did not change substantially, so that any changes in organisms were slow under the persistently uniform conditions.

The Lutetian transgression occurred in a province disturbed by folding, with the folds previously exposed to subaerial or submarine erosion; it also spread over the somewhat deformed continental massifs. Deposited over these massifs and the newly annexed regions were sediments with *Bulimes*[7] (Mt. Noire, and especially the northern part of Burgos Province). Littoral zones, particularly the principal ones, were the sites of deposition of thick limestone with corals, algae, large nummulites, assilinas, and alveolinas of the *Alveolina elongata* group (western Aquitania, Cantabrian coast, the northern part of Ebro land). The same limestone, apparently represented by thinner beds with intercalations of thin marl probably was deposited along the present axial zone where one of the more recent uplifts may have occurred on the site of interior massifs (Figure 3). It was connected with a continental plateau surrounding the Massif Central by way of the "Ariège shallows". Marl followed by flysch was deposited about that slightly submerged uplift. The marl carries thin intercalations of limestone with turritellas, assilinas, and nummulites (Aude, Gannes, Catalonia, Aragon, Navarre). The flysch is developed generally south of the uplift.

This structure deserves attention because it suggests the peculiar aspect of a process which originated in Lutetian time and determined the structural features of the Pyrenean province in the Tertiary. It was at this time that its axial zone underwent a minor subsidence and revealed the connection of its basement with the Massif Central, while the folded basement, represented in the south by Ebro land — appeared to split off the axial structure. A sinking trough of the Biscay-Catalonian strait was formed at the site of the northern segment of Ebro land. In the future, this trough served as a repository for the flysch facies.

This new distribution of land and sea in the Lutetian retained only three areas of active marine deposition, whose development took different courses. Without going into what happened after the *turritella* marl deposition,

[7]Terrestrial gastropods from the Helicidae family. Russian Editorial Board.

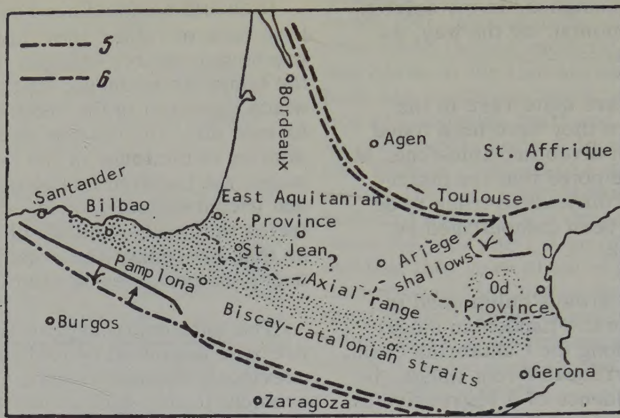

FIGURE 3. Paleogeographic map of the Paleogene after the Early Lutetian. After J.P. Mangin, 1959

1-4 - see Figure 2; 5 - assumed Cuisian coastline; 6 - ditto for the Lutetian.

we note that the Aude-Perpignan sedimentary province was filled up quickly and the sea retreated to the present Gulf of Lions. The development in western Aquitania proceeded at a slower rate; a slow westerly expansion of the Ariège isthmus, which involved the vast continental shallows, eventually restricted the area of marine sedimentation.

Finally, in the south, the downwarping of the bottom of the Biscay-Catalonian strait created an ideal province for flysch sedimentation; this province could extend as far as the axial zone. In the Lutetian, clastic flysch material came chiefly from Ebro land which has probably been uplifted in the south as a result of sharp subsidence in its northern part. At this point, sedimentary conditions were simple, considering the special nature of the sediments during continual tectonic movements with one side uplifted and the other depressed. The Ebro plain, a northern continuation of the bast Meseta, was a constant source of fine silty and pelitic material for the submerged province and its farthest reaches (the axial zone); this material could be dumped into the open sea by mighty flows of fresh water.

This appears to have been the cause of the great influx of terrigenous components brought in by fresh water, in suspensions granulometrically finer than usual. These suspensions proceeded to coagulate as soon as the continental and the marine waters mixed. The mixing of waters must have been very thorough for the elements to become mixed before deposition; for this reason, the grains within a layer remained unsorted.

The alternation of very fine-grained layers with predominantly pelitic particles, and layers with more silt particles is possibly due to climatic changes or, more likely, to a gradual change in the profile of equilibrium. When such an equilibrium has been established, only fine particles come in from the land, with coarser ones, up to gravel size, appearing with a revival of erosion. The climate over the land appears to have been very humid and probably warm, on the basis of the rapid rate of weathering. Such a climate promoted the growth of vegetation, which is reflected in the distribution of plant remains in the flysch rocks.

In many places, local rapid subsidence, faults, and development of horsts considerably disturbed the equilibrium of the sedimentary process; in these places, the terrigenous material was represented by conglomerate. However, the presence of conglomerate interbedded with flysch can be explained in another way, by the fact that when the rock disintegration process was less intensive or was slowed down by a very resistant limestone bed, fewer products of erosion were carried out of the source area. Finally, the conglomerate beds contain components of the "soft pebble" type, apparently separated from the sediment in the process of diagenesis and even during deposition itself. These pebbles are commonly irregular and emplaced in cement of the same composition. In such instances, we deal simply with redeposition of deep sediments, perhaps as the effect of a strong laminar flow (the tsunami type) which originated in connection with a sporadic intensification of deep-seated tectonic movements, either on land or in the subsidence area.

The flysch section carries isolated beds of nestone with alveolinas, discocyclinas, and ımmulites. They reflect interruptions in the flux of terrigenous material, having been de-›sited in shallow water. These beds occupy a rge area; thus, it turns out that almost all of e sedimentation province lay in shallow water. his fact, together with the absence of sorting flysch beds, suggests that we deal here with iallow-water sediments, on the whole more milar to those being deposited now on vast ›ntinental shallows which extend from a low ing land, rather than to those sediments buried ısorted in the deeps about cordilleras.

However, the Biscay-Catalonian strait was not lled up by flysch alone. Marly limestone with rritellas and corals are present in its central ea and in the east, developed as an extension the Aude limestones (Ribagorzane, Catalonia). onglomerate beds at the top of this sequence, uth of the present Maudit Mountains, suggest e presence of land areas in that zone.

The most typical Lutetian foraminifera fauna divided into two groups: large foraminifera epresented chiefly by large nummulites (group N. millecaput, and N. aturicus) accompanied / species known from the Cuisian stage and by ssilinas (A. granulosa, A. leymerii, A. ex-›nens). These forms are supplemented by ongated alvedinas (A. elongata, A. violae, A. gantea) and Orbitolites complanatus s. l. Dis-›cyclines are also well represented by Disco-/clina discus, D. archiaci, D. sella, D. num-ıulitica, and finally by Asterodiscus stellatus: perculina canalifera is present mostly at the ase. Finally, Eorupertia magna and Fabiania . cubensis occur without noticeable changes om the base to the top of this stage. These ›ssils abound in the limestone, with algae Lithothamnium album, L. nummuliticum) and ›rals often present in the lower intervals. ommon in marl and flysch are small pelagic ›rms: Globigerina trilocularis, G. (Catapsy-rax) dissimilis, G. linaperta, G. rohri, Porti-ılasphaera mexicana and Halkyardia minima. his assemblage of small species occurs in a alcareous facies. Finally, numerous ostracods re present in many beds of sandy marl in the avarre Superieure flysch.

Thus the distribution of structures is quite ell outlined in the Lutetian of this province. eep reaches of the Pyrenean strait, now sup-lied with sediments from its continental fringe nd destined to become the central axial zone, isplayed a tendency for forming an uplift, a range" connected by the Ariège isthmus to the ıargin of the Massif Central. At the same time, ıe axial Pyrenean zone split off Ebro land which egan to sink rapidly, creating a trough. The atter began to receive flysch under a shallow over of water. Limestone with nummulites, ssilinas, and alveolinas were deposited on levated stable segments where clastic material was missing. Marl, interbedded with the assilina and nummulite limestones, spread over depressed areas. This was the beginning of Tertiary movements which were to restore the Hercinian mountain structure.

By the close of the Lutetian, a brief period of equilibrium appears to have set in, between the subsidence which involved Ebro land, and the continued uplift in the axial zone. Under the shallow conditions prevailing in the Pyrenean branch of the sea, limestones were deposited nearly everywhere. These were limestones with giant nummulites including Fabiania and Eorupertia, the last of genera Assilina and Alveolina. Exposed more or less in those places where erosion has spared them, these limestones are known from Santander to Aquitania and in Navarre (Catalonia). It is quite probably that they also were deposited in the zone of axial highs; if so, no trace of them has been left.

The period of equilibrium in the basement movement was very important. It heralded the approaching end of the Pyrenean branch of the sea and accompanying momentous geographic changes. Indeed, from then on, the axial zone gradually emerged, simultaneously extending the dry surface of the Ariège isthmus, and pushing the sea farther west and south. On the Spanish side, however, the area of subsidence was extended progressively to the south, across Ebro land, as an effect of the central zone uplift.

The Pyrenees were in the making but the mountain formation was still a long way off (Figure 4).

And now we are at the base of the transition beds. This is no longer the Lutetian, nor is it quite the Upper Eocene. Assilinas and alveolinas have disappeared and the last of the large nummulites, Eorupteria and Fabiania, now have associated with them the first nummulites of the Nummulites striatus — N. contortus group. Discocyclinas are represented by D. strophiolata, D. varians, and Asterodiscus stellaris. This is the time of emergence of that part of the axial zone which from then on became the source of sediments. It was quite an unusual source province with much of its material belonging to the terrigenous group represented by fresh sediments, with some of them still in a diagenetic stage. On the other hand, some areas in the present zone of the Central Pyrenees, particularly east of the Malladette meridian, witnessed the erosion of sizable bodies of the Garumnien deposits which were plentiful here and not covered by marine Paleogene sediments. Such probably is the origin of the "tramo rojo" unit of Catalonia; it overlies the marine Lutetian and repeats in detail the Garumnien section, from the erosion of which it originated.

The emergence of the axial zone should be regarded as a long process contemporaneous with

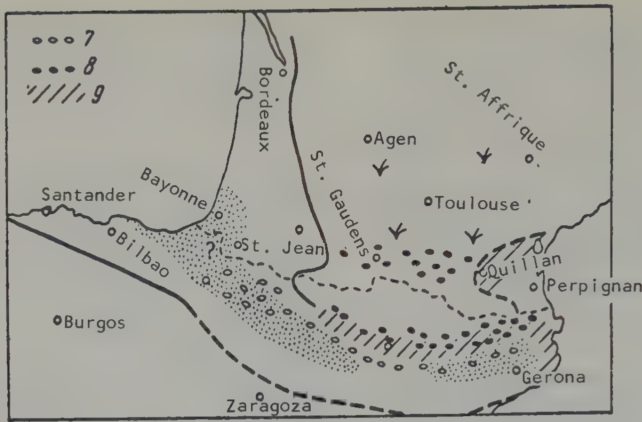

FIGURE 4. Paleogeographic map of the time of formation of the transition zone between the Upper and Lower Eocene. After J.P. Mangin, 1959.

1-6 - see Figures 2 and 3; 7 - marine conglomerate; 8 - conglomerate of subaqueous continental origin; 9 - slightly uplifted littoral areas.

sedimentation. This uplift was progressive in a deep basement corresponding to the ancient Hercinian structure, or else it was connected with a contemporaneous sinking of peripheral Lutetian structures in the axial zone and the Ariège isthmus. This process was terminated during the deposition of transitional beds, when a strip of land emerged from the sea within the future axial zone. The land strip immediately embraced the "Ariege isthmus" formed by more ancient elevated segments of the bottom and was fringed by lateral zones extending west and east and by the Biscay-Catalonian strait.

Terrigenous components accumulated in two sedimentary zones; in the north, the accumulation zone was located on land, within the new Ariège isthmus between Pau and Quillan. It received the first products of erosion of the axial dry land (Figure 5), including conglomerates not necessarily indicative of very broken up terrain; also molasse which came from the Massif Central, to judge from the composition of its heavy fraction. On both sides of the isthmus, the northern submerged province could maintain a marine regimen only in Roussillon (i. e., in the east of the Pyrenees) where marine conditions appear to have persisted up to the Pliocene. The west Aquitania zone, too, was being filled up with marine sediments. This stage is the Upper Lutetian of some authors, or the Auversian of Boussac, represented by marl and sandy limestone with Fabiania, Halkyardia minima, Discocyclina sella, D. pratti, etc., and the first nummulites of the N. striatus — N. contortus group (marl with pentacrynas from Biarritz, the vicinity of Orthez, etc.).

In the southern part of the Biscay-Catalonian strait, the only vestige of the Pyrenean sea branch, now pushed to the south, continued to sink under the weight of sediments continuing to arrive from the axial zone; these sediments formed flysch facies because they were deposited under the same conditions as before, in a sinking zone along the margin of a slowly rising land. The noticeable phenomena of redeposition of recently deposited Lutetian flysch are present in this flysch, especially in Navarre and Hujasco; the fauna of Lutetian globigerinas and globorotalia, formerly present, now are replaced by hardly distinguishable species (Globigerina stonei, G. Yequaensis, G. parva, and Globorotalia bullbrooki).

The local presence of conglomerate (aragon Superieur) suggests a continued uplift in the axial zone and a long erosion facilitated by the lack of consolidation in most of its component rocks. They became part of a low-lying land with isolated masses of Eocene and Cretaceous limestones and stumps of red Permian-Triassic rocks.

In the region south of the Maudit and Serdagny Mountains, the "tramo rojo" unit does not appear to have been deposited under marine conditions, because the axial zone uplift extended that far south. This uplift, contemporaneous with the sedimentation, undoubtedly was the cause of folding observed in this area, and of the unconformities at the base of the transition units.

Near the south side of the Biscay-Catalonian strait, the more or less considerable influx of terrigenous material points to the proximity of Ebro land from which it came. This material was deposited in a carbonate medium where it

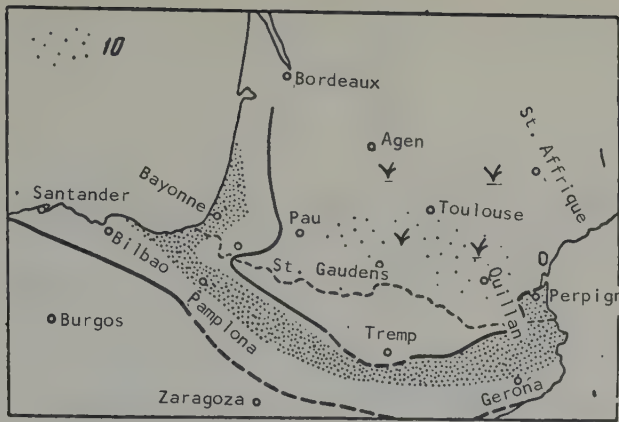

FIGURE 5. Paleogeographic map of the Late Eocene. After J.P. Mangin, 1959.

1-9 - see Figures 2-4; 10 - molasse deposits.

formed a thin bed of slaty sandy limestone with Fabiania, Asterodiscus and common Nummulites striatus-contortus, accompanied by very sandy marls with Porticulasphaera mexicana, numerous Halkyardia minima, and some benthonitic foraminifera associated with the Upper Eocene.

The time of deposition of this transitional unit occurred at the beginning of an unquestionably orogenic movement which initiated the Pyrenean range. The core of the axial zone, perhaps carrying on the ancient Hercinian structural plan of the Mt. Noire and the Massif Central, continued its slow rise and pushed the sea westward, eastward, and especially southward, on to the ancient sinking land mass. This controlled the sedimentary conditions. There is no doubt that most terrigenous material came to the Biscay-Catalonian strait from the north; it consisted chiefly of freshly deposited material.

The process so originated continued in the Late Eocene; throughout the Pyrenean province, it was characterized by the deposition of very fine-grained components, in both the continental molasse and "blue marl". Conglomerate beds, overlying the transition unit conglomerates, undoubtedly are the result of erosion of this mountain ridge.

The distribution of land in the Late Eocene differed little from that during the deposition of the transition unit (Figure 5). Dry land, developed out of the Ariège isthmus, extended from there to the west and east, while foraminiferal "blue marl" of the Pays des Basques was deposited apparently only beyond the Pau meridian. On the Ariège isthmus (Garonne Supérieure, Ariège, Aude), continental deposits continued to be deposited; they came from the Massif Central or the axial zone, which continued its uplift. As on the south slope, this sedimentation, contemporaneous with the uplift of the province, could have been accompanied by development of some unconformities.

The Biscay-Catalonian strait persisted in the south. However, the central zone uplift, still limited in width, slowly expanded to the south, especially in the Since-Ter watershed, and extended the land in that direction. In addition, the uplift was the most intensive right there, as witness the appearance of conglomerate. This structure was responsible for the peculiar submeridional swell, the "Ribagorzane shelf" (Figure 5), separating the two fairly active basins: The Navarre-Aragon in the west and the Catalonian (Olyana-Manlé) in the east. This brought about the formation of submeridional folds (i. e., the Ezer folds).

The Navarre-Aragon basin undoubtedly was connected with that of the Franco-Spanish Pays des Basques, because its marls are quite similar to those of Biarritz and the Catalonian basin, despite the presence of the "Ribagorzane shelf".

All this suggests uniformity in Upper Eocene deposition, in a sea which surrounded the axial zone.

The marine fauna present in the marl sequence developed slowly; it consisted mostly of such typical facies as Globigerina venezuelana, G. ouatchitaensis, Globorotalia centralis, Angulogerina halkyardi, Bolivina nobilis, Bulimina subtruncana, Marginulina behmi, and Nonion halkyardia. However, the number of Halkyardia minima increases toward the top of this sequence, heralding the future predominance of this typical form of the Oligocene. The coupled forms of Nummulites striatus-contortus gradually disappeared to give place to nummulites of an Oligocene aspect (N. fabianii and N.

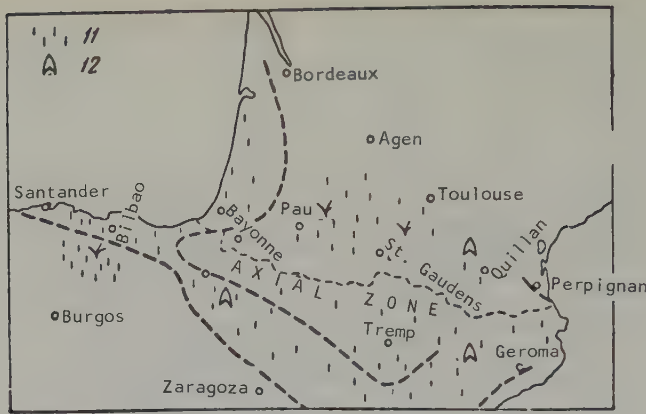

FIGURE 6. Paleogeographic map at the beginning of the Oligocene. After J.P. Mangin, 1959.

1-10 - see Figures 2-5; 11 - accumulation of clastic material; 12 - evaporites.

intermedius). Finally, littoral facies began to appear locally, even in the area of blue marl which is regarded as a deep water deposit.

The oneset of the Oligocene was marked by an almost general emergence throughout the Pyrenean province. To be sure, two marine zones persisted: one on the site of the present coast of the Franco-Spanish Pays des Basques, which fringed the Gulf de Gascogne; and the second, rapdily involved in the uplift, on the side of the Biscay-Catalonian strait. Early in the Oligocene, both zones were still connected by a strait between Navarre and Guypuscoa (Figure 6), which determined the similarity of facies in the Biarritz-Santander area, on one hand, and the Pamplona, on the other. In both areas, sandstone with Nummulites intermedius, Operculina alpina, and coral reefs were formed. Then the massif within the Pays des Basques[8] emerged, also in accord with its Hercinian structural plan; its area was added to that of the axial zone. Linked at depth with the Asturian massif, it raised its own sedimentary mantle above sea level, thereby finally severing the connection between the Atlantic and the Mediterranean basins.

The marine province thus divided was considerably restricted, so that while marine deposits on the Pays des Basques coast persisted extensively until the Late Oligocene, gypsum appeared in the now isolated Navarre-Aragon basin, initiating the deposition of a fairly complete halide sequence (potassium salts of Navarre). The Catalonian basin, less isolated from the Mesogea[9] and less encumbered by disintegration products of the surrounding highlands, was perhaps the site of maximum accumulation (Cardona evaporites).

Coming next are thick molasse beds; they filled up the former Biscay-Catalonian strait and produced the present Ebro basin. There are two conglomerate sequences wedging into the 3000 to 4000 m thick silt deposits, commonly with a cyclic structure. The conglomerate members, of variable thickness, indicate the distribution of river courses and suggest the regional scope and sporadic intensification of the erosion process, on the premise that when the distribution of conglomerate changes in space, it is narrowly restricted in time.

North of the ever-rising central zone, the Oligocene sedimentation was almost fully continental, except for embayments in the Pays des Basques and perhaps the Perpignan region (Figure 7).

On the Spanish side, the age of the thick continental molasse sequences has been determined, in the absence of a fauna, from their relationship with overlying and underlying deposits. Present at the base are the first Oligocene foraminifera, while the top, generally eroded away, is overlain unconformably by continental deposits with the familiar abundant fauna of fairly high Miocene beds. Consequently

[8]In the western part of the Pyreness, south of Biarritz. The Russian Editorial Board.

[9]The Mediterranean basin. Russian Editorial Board.

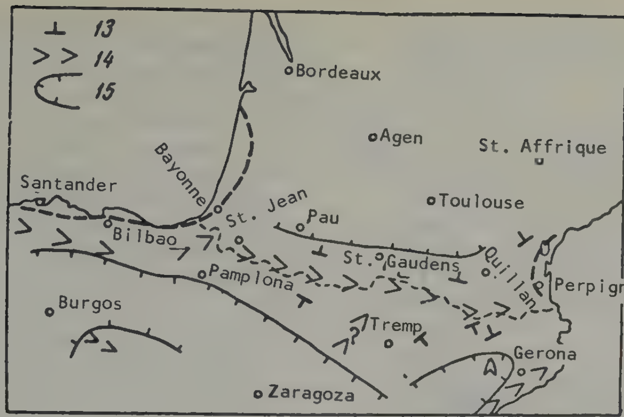

FIGURE 7. Paleogeographic map at the end of the Oligocene (or beginning of the Miocene). After J.P. Mangin, 1959.

1-12 - See Figures 2-6; 13 - direction of overturning of main folds; 14) axial zones of mountain structures; 15 - continental depressions with folded basements, filled up by the erosion products.

these erosion products, molasse and conglomerate, are already Oligocene, although it is impossible to determine whether they represent all of the Oligocene or only a part of it.

The age of the continental clastic sequences is just as difficult to determine on the French side. M. Casteras (1933), page 31) admits that "to tell the truth, there are no certain data in the eastern part of the Pyrenees on the precise age of the main paroxysm. It is known only that a definite change in conditions took place in the Late Lutetian. The nummulite sequences are followed by puddings (conglomerate) of Palassou. However, no unconformity has been observed in the entire section, in the Carcassonne basin, between the Pyrenees, Moutoumé, and Mt. Noire. In the Garonne Superieure, on the south flank of the Aquitanian basin, Miocene molasse rests unconformably on the Palassou conglomerate. Thus, a very wide gap remains in determining the age of the main folding".

With this thick molasse, Paleogene deposition in the Pyrenees came to an end.

To recapitulate: a slight uplift, trending apparently west-northwest to east-southeast, slightly deformed the older deposits in Early Lutetian time. The Lutetian itself is transgressive and locally unconformable. Beginning with the close of the Lutetian, and proceeding as an extension of the uplifts, a mountain chain was outlined, with its axial zone slowly rising. Local thrust stresses were operative from the direction of the southern submerged province, comparatively deeper than the Ariège isthmus, along the periphery of the central uplift (Aragon, Catalonia), while the sedimentation around it proceeded undisturbed.

It is difficult to visualize the importance of paroxysms responsible for the mountain chain and separated by quiescent phases, in the overall course of histotical development. As to the formation of local folds in the immediate vicinity of the axial zone, there was no general folding prior to the termination of deposition of clastic Oligocene sequences, inasmuch as these deposits rest mostly conformably on older units with which they are folded and involved in major thrusts.

Undoubtedly, the Pyrenean range was formed in many stages closely related in time. There is evidence to the effect that the folding is younger than the cores of this mountain chain. These uplifts appear to have occurred at different times in different places. Local differences can be determined, and it can be assumed that their tectonic pattern was determined by the mosaic structure of the deep basement which consisted of relatively independent elements but connected perhaps by very ancient roots to the Massif Central of France.

I believe that the following major conclusion is warranted: the end of this long history is expressed in the appearance of a folded mountain structure, more or less overturned toward its foothills; the latter, too, are plicated, with the folds dating back to the Miocene, or the close of the Oligocene at the earliest.

THE "PYRENEAN MOUNTAINS", A POST-PALEOGENE FOLDED STRUCTURE

If a prolonged evolution of the deep basement has led to a horst uplift of block segments in the axial zone, the folding of the sedimentary mantle did not begin on the whole until the elevation difference between this horst and the peripheral zones of subsidence was sufficient for a break-off and sliding down of the numerous argillaceous-marly beds and flysch. From that time on, the mantle of axial uplift, and then perhaps the Hercinian core itself, contributed the products of their erosion to the marginal zones. Toward the close of the Oligocene or perhaps in the Miocene, this process was complicated by a general folding, different in distribution from the local folding preceding it, and brought about by the development of uplifts. The sedimentary mantle broke off under the action of gravity and slid in the direction of the foothills.

This movement could be accelerated by the intensified disturbance of equilibrium brought about by erosion which unloaded the axial zone and provided sedimentary material, thus promoting a further subsidence of the marginal zones. In addition (this is purely hypothetical), the final push could have been provided by an abrupt break between the axial zone and the Ariège basement. This possibly was a rejuvenation of an ancient break as the effect of tilting of deep segments of the axial zone because of the subsidence of a portion of Ebro land, in the south. The "north Pyrenean front mantle" originated or was revived probably at that time. The sedimentary mantle here started to slide northward and was sharply distorted so that it formed only a narrow mountain range. In the meantime, the southern mantle formed gentler folds (involving those formed in the period of formation of transition beds, as a result of spreading of the uplift), because it bordered sediments of immense thickness filling up the new Ebro trough; in this process, the southern mantle was disturbed in the lower part of the gentle slope, thereby defining the outlines of the Pyrenean Sierra foothills.

The mantle structures can be divided into three types. At the crest, the already attenuated mantle was stretched out; fairly regular folds were formed along the slope, becoming more compressed and more numerous down the slope. At the base of the slope, they are strongly disturbed and thrust over the foothills. The latter were deformed in the reaction, as witness the Ebro basin; however, the pressure was damped by the sedimentary section, giving rise only to broad undulating flexures. The latter enabled the Oligocene gypsum to cut through them in purely local vertical diapirs.

In western Aquitania, on the other hand, the sedimentary mantle of the foothills is thin and broken off at the Triassic level; here the pressure from the Pyrenees brought about the formation of several crests. The Triassic in these uplifts, parallel to the trend of the range itself, was locally "squeezed out". In addition, the numerous basement breaks in the relatively rigid Aquitanian Platform may be associated with the great North Pyrenean fault, and their arrangement may have determined the location of the Pyrenean front folds. In Languedoc, smooth folds in the sedimentary mantle are reminiscent of the Ebro basin undulations. Finally, along the Mt. Noire periphery and in Provence, the reaction to Pyrenean movements was more intensive in a thinner and more consolidated medium (the Stampian paroxysm of Provence, noted by G. Corroy and G. Denisot, in 1935 and 1942 — 5).

Thus the Pyrenean province originated on the whole after preliminary local movements, beginning approximately at the end of the Lutetian. It was involved in a general movement since about the close of the Oligocene, at the latest before the deposition of Pontian[10] beds. It is difficult to conceive that there was a long time break between the formation of "mountain range folds" and "plainland folds". In the Pays des Basques, for instance, the axial zone uplift caused a displacement in the Oligocene mantle, so that folds caused by this uplift in Pays des Basques and Navarre were arranged with the same parallelism, as far as the central Pyrenees foothills. In addition, the shocks of these movements, transmitted by way of meridional faults, have affected the youngest land regions, uplifted in the Oligocene.

From Asturia (united with the Pyrenees, as we have seen, by both deep-seated and surface ties) and to Provence, the mountain structure was formed and folded prior to assuming its modern aspect which came into being only in the Pliocene and Quaternary, after new transformations and erosion.

Even if the terminal folding stage did involve the entire Pyrenean province as a whole, it is unquestionably true that it manifested itself in a region of variable structure, where local folds and unconformities could have originated since the close of the Lutetian. Moreover, it represented the last event in a long sequence of fruitless efforts to restore the Hercinian structure, because, in the final reckoning, this Hercinian structure transpires fairly distinctly in the system of faults in the mantle, as early as the Cretaceous and again in the Paleocene; its undulations were more or less subordinated to the west-northwest to east-southeast trend of the short-lived Pyrenean strait. It is its block structure, perhaps of a very ancient origin, that has determined the structure of the sea bottom

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[10]Western European Pontian (terminal Miocene) is meant rather than the Russian Pontian (basal Pliocene). Russian Editorial Board.

nd the land, and the course of sedimentation self. A revival of uplift in individual blocks, nore or less united within the mountain range, as determined their tectonic aspect as well as ne time of formation of local folds. This Herinian structure transmitted its own features to ne axis of the long Pyrenean chain, to produce broken line fringing the Massif Central of rance. Finally, a mighty and simultaneous plift of this structure brought about the terminal general folding responsible for the present spect of the Pyrenees.

It appears, then, that the known facts, if not heir interpretation, are in accord with the ypothesis briefly set forth in this paper acounting for the peculiar structural features of he Pyrenean range. This history of the Pyenees, however brief and devoid of details hich could illustrate and strengthen it, may erve as the point of departure for further inestigations which may either corroborate or efute it.

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THE CHERKESSK-KELASURI TRANSVERSE DISLOCATION IN THE STRUCTURE OF THE GREATER CAUCASUS[1]

by

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The attention of students of the Caucasus has been attracted by the importance of meridional and northeasterly trends in its structure, transverse to the principal structural zones, which trend from northwest-southeast to east-west.

* * * * *

N. S. Shatskiy was first to recognize the independence of transverse ("anticaucasian") structural trends. He has demonstrated that the tectonic development of the Greater Caucasus is determined to a considerable extent by a major meridional flexure involving not only the Caucasian Range and the Cis-Caucasus but the adjacent parts of the Russian Platform, as well [11].

The uplifted limb of this flexure in the present structure of the Cis-Caucasus is the Stavropol uplift; its continuation within the Caucasian Range has brought to the surface large areas of ancient Paleozoic rocks. Subsequently, many students noted smaller "anticaucasian" structures in various areas of the Greater Caucasus; some of these structures are faults, others are folds and flexures [5, 8, 9].

The interest of Caucasian geologists in these transverse structures has also been heightened by the fact that associated with them are the largest known Caucasian ore deposits.

This paper deals with one such structure which cuts across the Caucasian range approximately along the Cherkessk-Sukhumi line (Figure 1). We have named it the Cherkessk-Kelasuri structure.

This structure crosses various tectonic zones of the Greater Caucasus; among them, the following are identified on our map: 1) the Main Range zone of Lower Paleozoic metamorphic and crystalline schists with Caledonian intrusions and an immense body of Hercinian granite; 2) zone of the north slope of the Main Range, most of which is covered by Mesozoic and Cenozoic deposits dipping monoclinally to the north; only in the south, in the Peredovoy (Front) Range, Paleozoic formations come up from under them; 3) the south slope of the Main Range with its complexly dislocated Mesozoic and Cenozoic deposits.

Inasmuch as these zones differ in geologic structure, the Cherkessk-Kelasuri transverse trend is expressed in formations of different ages. We shall trace its effect in all three zones.

In the Teberda-Aksauta watershed, the structure of Paleozoic rocks exposed in the Peredovoy Range zone has been complicated by a major meridional fault. West of it there is a large area of the Aksauta transverse trough with upper Paleozoic deposits. East of the fault, the upper Paleozoic has been preserved only in narrow longitudinal grabens, with Devonian and Lower Carboniferous deposits exposed (the Gidam uplift). The displacement along this fault, in the northern subzone of the Peredovoy Range[2] is no less than 800 m; in the main and southern subzones, the fault does not disturb the continuity of upper Paleozoic rocks, being expressed only in an abrupt easterly rise of the flexures and in the appearance of submeridional trends (the Gidam-Otar Range).

A study of the history of development of this region shows that the Aksaut trough and the Gidam uplift, tectonically different at the present time, underwent differential movements, beginning during the Middle Carboniferous and lasting through all of the late Paleozoic. These movements were reflected in the shift in the

[1] O Cherkessko-Kelasurskom poperechnom narushenii struktury Bol'shogo Kavkaza.

[2] V. N. Robinson identifies the northern, main, and southern subzone in this area of the Peredovoy Range.

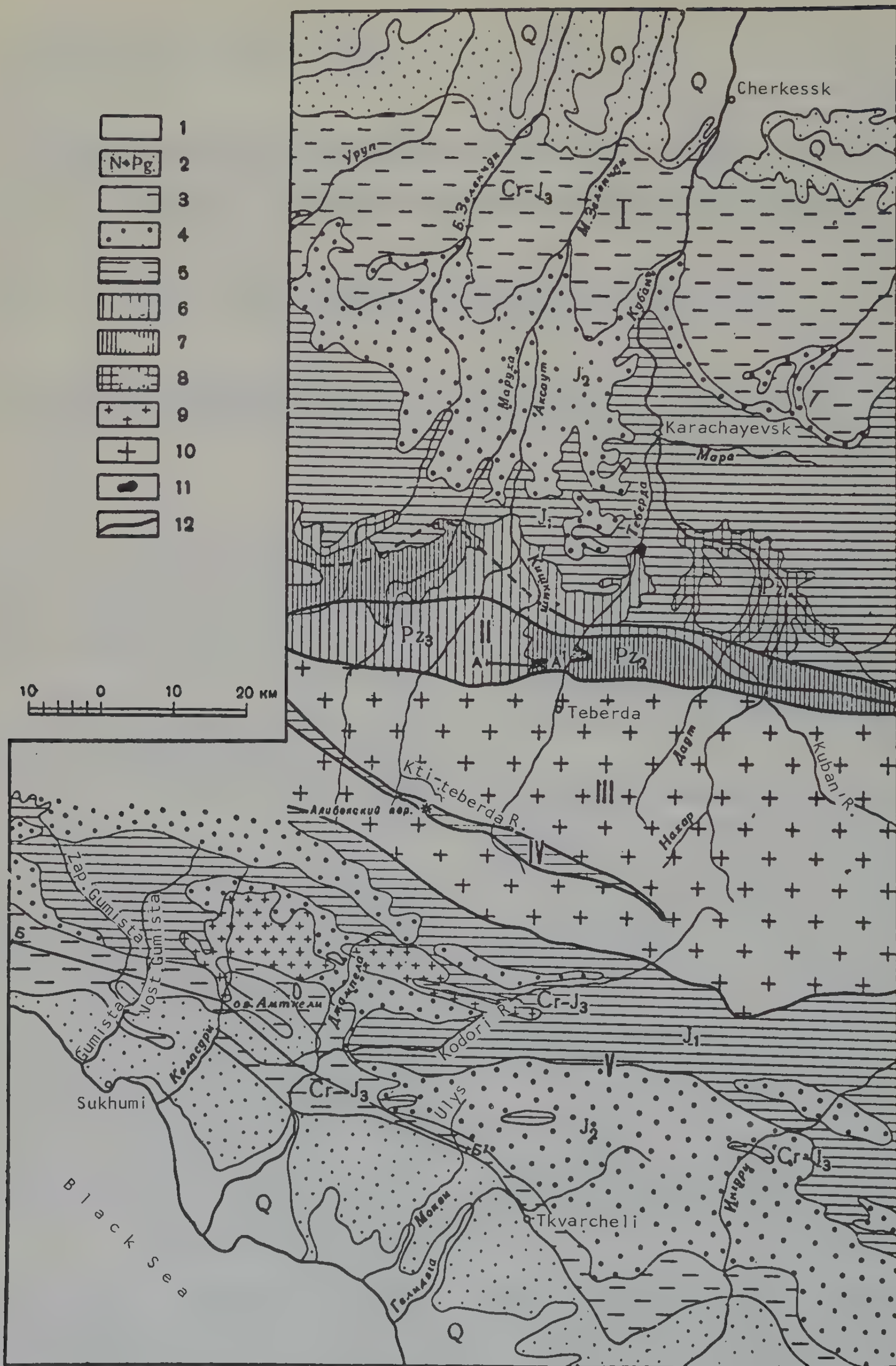

FIGURE 1. Generalized geologic map of the area of the Cherkessk-Kelasuri transverse dislocation.

1 - Quaternary deposits; 2 - Tertiary deposits; 3 - Cretaceous and Upper Jurassic deposits; 4 - Middle Jurassic deposits; 5 - Lower Jurassic deposits; 6 - upper Paleozoic; 7 - middle Paleozoic; 8 - lower Paleozoic; 9 - Jurassic granite; 10 - Paleozoic crystalline schist and granite; 11 - serpentinite; 12 - fault traces.
I - the north slope monocline; II - the Peredovoy Range zone; III - the Main Range zone; IV - the Arkhyz-Klych trough; V - the south slope zone.

boundary between the eroded land and the sedimentation area, in late Paleozoic time, as well as in the distribution of facies and thicknesses of upper Paleoozic sediments. It was the Gidam uplift that represented the uplifted land source of upper Paleozoic clastic material for the adjacent sedimentation area coinciding with the Aksaut trough. In that trough, the thickness of all upper Paleozoic sediments decreases west to east, toward the Gidam uplift (Figure 2). Thus, the Middle Carboniferous in the south subzone of the Peredovoy range is 600 m in the axial part of this trough and zero on the west slope of the Gidam uplift. The Upper Carboniferous thins out in the same direction, from 700 m to 400 m in the north subzone and from 800 m to nothing in the south subzone. The thickness of the Lower Permian Aksauta formation ranges from 815 m to nothing, and the upper red formation from 5000 m to 400 m in the south subzone, and from 1600 m to 1200 m in the north subzone. A coarsening of sediments has also been observed in the Aksauta formation, from the west toward the Gidam uplift, while the upper red formation contains an increasing amount of middle Paleozoic gravel.

North of there, in the north slope monocline area, the Cherkessk-Kelasuri dislocation is fixed by the position of the Pliocene peneplain. According to V.I. Fomin, east of the Kuban this peneplain is located at an elevation of 2320 m (The Bichesynsk Plateau), but it lies considerably lower, at 1500 m, to the west. This means that the area east of the Teberda was uplifted 700 to 800 m, relative to the western area, in post-Pliocene time.

A study of the geologic map of the monoclinal north slope of the Main Ridge reveals that the Bichesynsk Plateau and the area west of it have different structures. Where both Lower and Middle Jurassic deposits are widely developed west of the Teberda and Kuban rivers, the Middle Jurassic is missing in the Bichesynsk Plateau east of these rivers.

Thanks to the preserved Oligocene peneplain, it may be assumed that differences in the geologic structures of the two areas separated by the Teberda and Kuban rivers are due to the fact that the eastern segment was already uplifted relative to the western, during the Pliocene and prior to peneplanation. Considering that there is no appreciable coarsening in Middle Jurassic deposits of the western segment toward the Teberda and Kuban, and no thinning, it must be assumed that the absence of the Middle Jurassic on the Bichesyansk Plateau is related to post-Middle Jurassic movements. Thus, the two adjacent segments underwent a differential movement as early as somewhere between the Middle Jurassic and the Pliocene.

In addition, an analysis of the occurrence of the Lower Jurassic in the Teberda-Kuban basin shows that Lower Jurassic deposits east of these rivers rest mostly on the lower Paleozoic, and

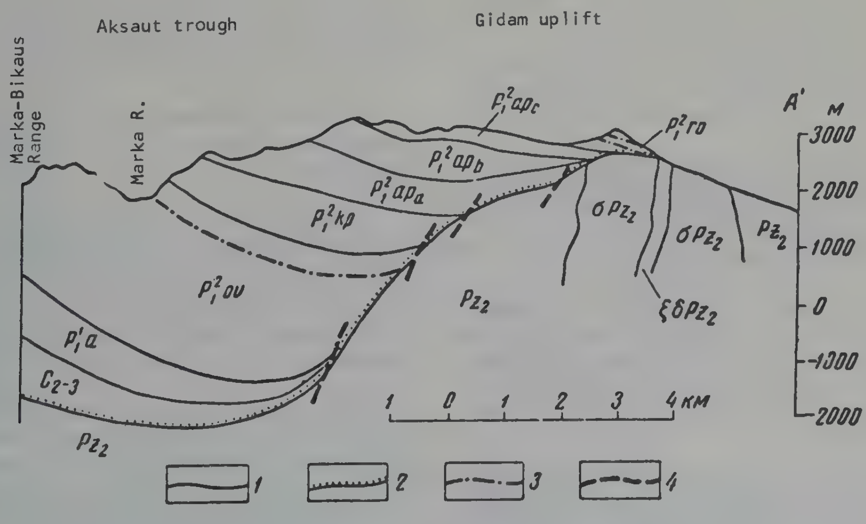

FIGURE 2. Cross-section along the Peredovoy Range south subzone.

A - A[1]. No vertical exaggeration. Upper red formation $P^2{}_1$ ro, a sequence of rhythmically alternating conglomerate, sandstone, and siltstone; silty-sandy subformation $P^2{}_1$ ro_c, motley siltstone, sandstone, and shale; $P^2{}_1$ kp, conglomeratic sandstone subformation; $P^2{}_1$ ov, volcanoclastic subformation; $P^1{}_1$ a, the Aksauta formation; C_{2-3}, Middle and Upper Carboniferous; Pz_2, middle Paleozoic (Devonian - Lower Carboniferous); σ Pz_2, serpentinite; $\xi\delta Pz_2$, syenite-diorite; 1 - boundaries of conformable and intrusive contacts; 2 - stratigraphically unconformable contacts; 3 - individual beds; 4 - faults.

on the middle and upper Paleozoic west of there. Likewise, the occurrence of upper Paleozoic deposits is different on either side of the Teberda and Kuban. East of them (in the north subzone of the Peredovoy Range), upper Paleozoic rocks, not well developed here, rest upon the lower Paleozoic; west of them, they rest on the middle Paleozoic.

Thus differential vertical movements of the two adjacent segments of the crust along the Cherkessk-Kelasuri transverse dislocation occurred in the north slope monocline, from at least pre-late Paleozoic time to the post-Pliocene. As in the Peredovoy Range province, the east segment was consistently higher than the western.

Another example of the movement along the Cherkessk-Kelasuri dislocation in the north slope zone of the Main Range is a meridionally trending thrust in the area of Nizhne-Taberdinskoye village, described by G. D. Afanas'yev [1]. Here, lower Paleozoic metamorphic schists have been thrust from the east over Lower Carboniferous deposits.

The intensification in igneous activity of different ages along the same line is noteworthy, in the north slope monoclinal zone. This is true for middle to upper Paleozoic porphyrite, Lower Permian albitophyre (keratophyre), and Jurassic andesite-dacite.

Judging by our data, the Cherkessk-Kelasuri dislocation in the south clope zone of the Main Range was first manifested during the Early Jurassic and particularly so in the Late Jurassic, Cretaceous, and Paleogene. It is also reflected in the present-day structure of the south slope.

In the Liassic (Triassic), as noted by I. R. Kakhadze [4], an uplift occurred in the Kelasuri granite massif area; the products of its erosion have been observed in sections along Bol'shaya Gumista and Kelasuri rivers.

In Late Jurassic time, a transverse flexure existed along the Cherkessk-Kelasuri line in the south slope zone of the Main Range; it affected not only the thickness of the accumulated sediments but to a considerable extent their facies changes as well.

Thus, where comparatively thick (over 1000 m) fossiliferous Upper Jurassic marine sediments, are present west of this transverse flexure, the Upper Jurassic east of there is represented by Upper Oxfordian and Lusitanian motley lagunal-continental deposits [6]. Consequently, a barrier was present along this flexure, which constituted the east boundary of a Late Jurassic marine basin.

The Cherkessk-Kelasuri transverse dislocation was even better expressed in the Early Cretaceous. A detailed study of Lower Cretaceous deposits in the Sukhumi area [3] shows tha a well-defined uplift existed along the Cherkessk Kelasuri line, during their deposition. That uplifted divided the Abkhaziya sedimentary basin into two troughs: the Gudaur in the west and the east Abkhaziya in the east. This transverse structure was steadily rising during all of the Early Cretaceous, which resulted in a sharp thinning of the sediments over it and even in a temporary break in sedimentation (Figure 3).

During the Valanginian-Hauterivian, this uplifted area was subsequently submerged in the Barremian and Aptian. A new intensive uplift occurred at the onset of the Albian. As a result of submarine and subaerial erosion, Lower Albian, Aptian, and some Upper Barremian deposits were obliterated. Subsequent redeposition resulted in the formation of a bed of limestone fragments, phosphate nodules, and fossils of Barremian, Aptian, and Lower Albian ages. Resting on this "interruption" bed are shale and marl with a Middle- to Late Albian fauna.

By the close of the Cenomanian, this uplifted area again stood dry, with continental red shale deposited on it. They are transgressively overlain by Turonian limestone with a thin basal conglomerate.

The presence of the Cherkessk-Kelasuri dislocation in the south slope zone of the Main Range is fixed also in Paleogene deposits. In the Paleogene, another relatively uplifted area was located along this line, with sedimentary breaks occurring between the Lower and Upper Paleocene and in the middle of the Middle Eocene as first noted by M. S. Shvetsov [10].

The present Sukhumi area, too, is uplifted, separating the trough of the Colchis plain from the Gudaut trough.

Associated with the Cherkessk-Kelasuri transverse dislocation is a major Jurassic intrusion, the Kelasuri granite massif. It is possible that this coincidence is not accidental and that the intrusion of the Kelasuri granite was due to a considerable extent to the presence of the Cherkessk-Kelasuri dislocation.

In the Main Range zone, the Cherkessk-Kelasuri line is marked by a sharp flexure in the structure of the Arkhyz-Klych Jurrassic trough which trends northwest among uplifted massifs of the ancient Paleozoic basement of the Greater Caucasus. West and east of the right bank of the Aksauta, along the line of that flexure, parts of the Jurassic trough exposed, are different in depth and structure.

The present Arkhyz-Klych Jurassic trough is a complex graben whose over-all structure is represented in Figure 4. West of the right flank of the Aksaut Valley, only the higher portions of the trough are exposed, with its much

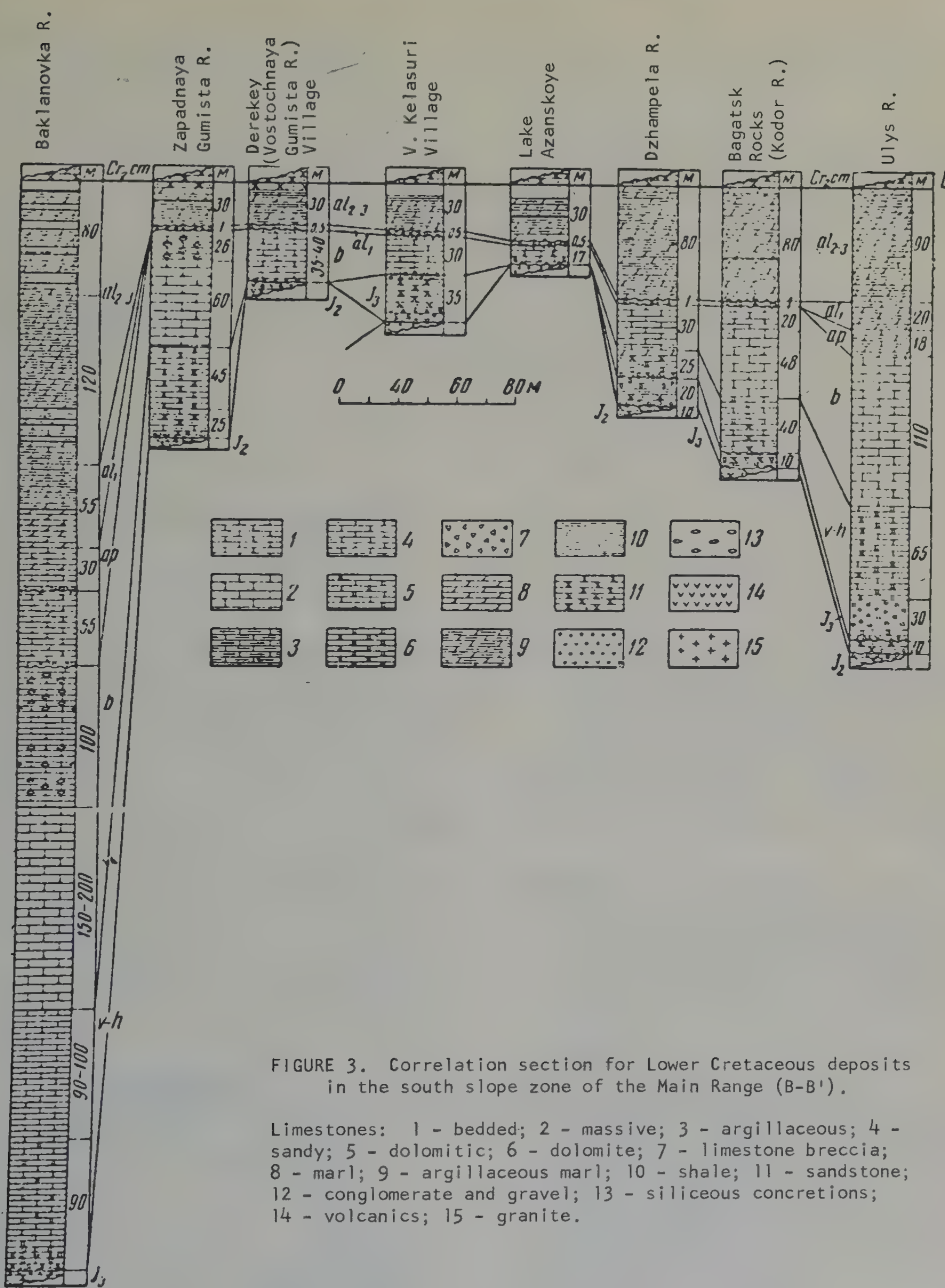

FIGURE 3. Correlation section for Lower Cretaceous deposits in the south slope zone of the Main Range (B-B').

Limestones: 1 - bedded; 2 - massive; 3 - argillaceous; 4 - sandy; 5 - dolomitic; 6 - dolomite; 7 - limestone breccia; 8 - marl; 9 - argillaceous marl; 10 - shale; 11 - sandstone; 12 - conglomerate and gravel; 13 - siliceous concretions; 14 - volcanics; 15 - granite.

leeper parts exposed to the east. Because of hat, the Arkhyz-Klych trough can be divided structurally into two parts, the eastern and the vestern; the dividing boundary passes along the ight bank of the Aksaut. Possibly because of his transverse flexure, the Arkhyz-Klych rough wedges out completely in the Aksaut valey; instead, its left side exhibits a number of breaks oriented obliquely relative to the trough.

Inasmuch as deposits younger than the Lower Jurassic of the Arkhyz-Klych trough itself are missing in the Main Range zone, it is difficult to say anything definite on the age of the transverse flexure in the trough structure. In any event, the formation of this flexure was completed in post-Early Jurassic time.

In the present-day relief of the Main Range, too, a transverse flexure is clearly defined along the Cherkessk-Kelasuri dislocation, north of the Jurassic trough. West of this line, absolute

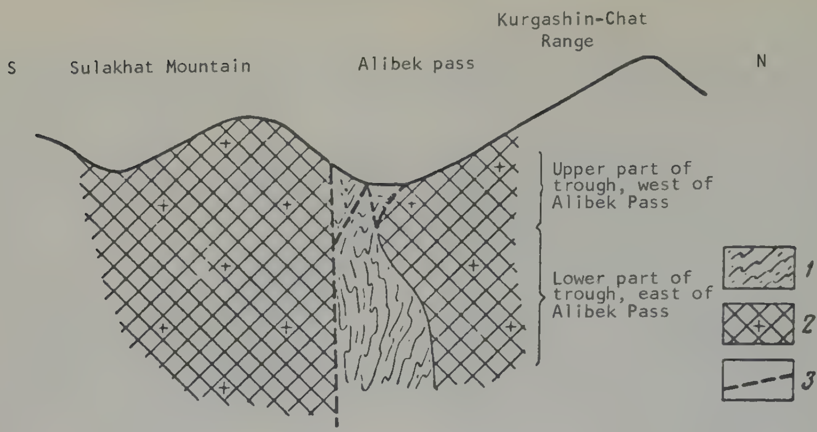

FIGURE 4. Diagrammatic cross-section of the Arkhyz-Klych trough.

1 - Lower Jurassic schist; 2 - Paleozoic crystallines; 3 - faults.

elevations do not exceed 2700 m, while east of it there are such peaks as the Semenov-Bashi, Musatcheri, and others, all standing above 3000 m. It is not impossible that the sharp easterly rise of the Arkhyz-Klych trough we have mentioned before is connected with very recent Quaternary movements which have controlled the present mountain landscape of the Main Range.

In connection with this transverse disturbance, it is of interest to consider the origin of the structure of the Kti-Teberda scheelite-arsenopyrite ore deposit on the right bank of the Aksaut, somewhat north of the Arkhyz-Klych trough.

The structure of this deposit was first described by A.V. Pek and L.I. Lukin [7]. They believe that crystalline and metamorphic schists of the Kurgashin-Chat Range form the south limb of a large anticline of the same trend as the Jurassic trough (N 60° W). The core of this anticline, exposed in the Kti-Teberda valley, is an intrusion of gray dimicaceous granite which has a normal intrusive contact with the enclosing rocks. The ancient schists, and the over-all structure, trend nearly latitudinally. However, northeasterly strikes were also observed by A.V. Pek and L.I. Lukin in lower Paleozoic schist of this area. In addition, these students, working with the orientation of quartz and mica in crystalline schist, recognized the presence of B-tectonite with a transverse "anti-caucasian" orientation of the B-axis (N 19° E), unusual for the Greater Caucasus.

In search for an explanation of this position of the B-tectonite axis, those authors reject the presence of a transverse folded structure and explain this phenomenon by a deformation of the enclosing crystalline schist due to horizontal flow of a viscous granite magma, from east to west. Their thesis is that the first explanation does not account for the nearly horizontal position of shearing plane S, as determined from mica, and the nearly vertical dip of meridional veins in the deposit; nor does it "find a confirmation in the over-all structure of the Caucasus.

A 1957 study of the structure of the Kti-Teberda deposit confirmed the B-tectonite structure in crystalline schist and gneiss. It was also determined that the maxima of the quartz optic axes, determined in a circular belt, generally correspond to the steep shearing plane S (Figure 5).

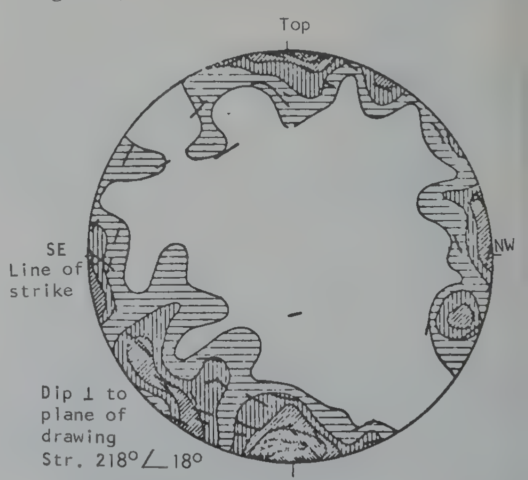

FIGURE 5. Diagram of quartz orientation in crystalline schists.

200 optic axes; isolines 1 - 2 - 3 - 4 - 5%. The paths of shifting of optic axes in quartz grains are indicated.

Consequently, one of the reasons for the velopment of submeridional veins in this posit is assumed to be primarymicrofractur-g, coinciding with the vertical position of the plane (in quartz) and apparently related to e formation of some transverse structure. was also demonstrated that intrusion of the i-Teberda granite had been accompanied by change in the microorientation of quartz, in ntact segments, corresponding to the flow of agma toward the crest of an anticline formed granite. Consequently, the intrusion of gran-e did not lead to the transverse orientation of e microstructure.

Similar results were achieved by a study of nphibolites which form thick (up to 25 m) lenses crystalline schist and gneiss. It has been stablished that the excellent axial schistosity amphibolite, corresponding to the orientation axis B of deformation, has a consistent sub-eridional (N 10° E) trend, in different parts the deposit, with a gentle dip (up to 10°) ther to the south or to the north. Unquestion-bly, this perfect orientation of amphibolite ould not have been produced by granite flow it was related to a long period of deformation ccompanied by recrystallization of the amphi-ole minerals. Small, in places isoclinal, folds hose axes trend northeast (N 25° E) are also resent in the gneiss.

Thus, a study of structure and microstructure f this crystalline section shows quite convinc-ngly that it has undergone a deformation whose -axis had a direction normal to the trend of rincipal Caucasian structures.

Such a study of ancient rocks in a limited rea with a comparatively undisturbed mono-linal occurrence of crystalline schist does not rovide a positive answer to the problem of his unusual orientation of the B axis. However, he concept of a transverse flexure complicating he crystalline basement is in fair accord with ata extant on the structure of this deposit; as uch, it can be accepted as an explanation for his phenomenon.

Petrotectonic studies have shown that quartz rains in dimicaceous granite exposed in the ore of the Kti-Teberda structure are not uni-ormly oriented. It can be assumed therefore hat the transverse flexure which complicates he over-all northwesterly-trending structure vas formed prior to the intrusion of Hercinian ranite.

SUMMARY

1. A major transverse dislocation is present long the Cherkessk-Sukhumi line; it cuts across he trend of major geotectonic zones of the Greater Caucasus.

2. This dislocation most likely originated prior to the intrusion of Hercinian granite, and was active up to the Quaternary.

3. The long life of this disturbance leads to the belief that it reflects on the whole a deep-seated fault which was expressed in different ways in different periods and in different geo-tectonic zones of the Greater Caucasus.

The identification of this transverse feature is of definite practical importance. Associated with it in the Main Range zone is the Kti-Teberda scheelite-arsenopyrite deposit. In the south slope zone of the Main Range, a sharp thinning and wedging out of individual Mesozoic units tools place along it. This last circumstance, in the presence of other favorable factors, may lead to the accumulation of oil and gas.

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CLAY MINERALS IN SEDIMENTS OF THE EASTERN PART OF THE BLACK SEA[1]

by

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This paper is part of a joint study of present-ay sediments in the Black Sea, being carried n by the Oceanography Institute, the Laboratory f Geologic Problems, and the Geological Institute, AS U.S.S.R., under the general direction f N.M. Strakhov.

The purpose of this study was to determine e composition of clay minerals in the eastern art of the Black Sea; the effect of diagenetic rocesses on the clay components at early stages f sedimentation; and the nature of changes in inerals with relation to the physiochemical nvironment. With this in mind, a detailed tudy of cores of the Black Sea oozes, taken long three lines has been carried out; opposite e mouths of Rion and Chorokh Rivers, and pposite Lazarevskaya village, a short distance orth of Sochi.

Along each profile, cores were taken from ifferent depths, 50 to 2000 m, at a distance of 0 to 90 km off shore. The average length of ores was 2 to 3 m, with some cores reaching to 9 m. Coring for a detailed laboratory tudy was done with consideration given to the epth of the sea, the depth below the bottom urface, and to petrographic and stratigraphic riteria.

The detailed study of mineral composition of e colloid fraction of the oozes was preceded y a macroscopic study, as a result of which aree stratigraphic units were described, coresponding to the Recent, ancient Chernomorian, nd Neo-Euxinian deposits identified by N. Strahov in the Black Sea deposits (Figure 1).

The Black Sea deposits are subdivided into wo natural groups: shallow-water sediments f the zone of oxidation and deep-water sediments of the hydrogen sulfide zone. Paleontoogic criteria have been adopted in differentiaing shallow-water deposits; lithologic criteria re used for deep deposits devoid of any faunal emains.

The sediments under study are located along the Caucasian coast whose specific structural and relief features are reflected in some specific features of its sediments, such as a high silt content in oozes and the scarcity of organic matter in shallow-water sediments. Recent sediments have been strongly affected by Rion and Chorokh Rivers which provide large volumes of terrigenous material. Deposits located directly off the mouths of these rivers, unlike the typical shallow-water deposits, are utterly devoid of faunal remains and are represented by comparatively coarse silty-argillaceous material with well-defined stratification brought about by differences in the mechanical composition of individual layers as well as color and the organic matter and iron-oxide content [6, 7, 11].

Two varieties have been identified among Recent sediments of Group Three, the oxidation zone: littoral deposits (Station 18) and shallow-water Mytilus ooze (Station 17).

Littoral deposits are represented by coarse silty argillaceous material with an abundant and diversified fauna typical of the S. A. Zernov littoral sand facies. The predominant form here is Gouldia minima, with Syndesmia ovata, Nassa reticulata, and Cardium mactra common, and Mitylus, Caliptera, Modiola, Tapes, worm tracks, and isolated bryozoan colonies less common.

The Mytilus ooze is a gray homogeneous clay with rare intercalations of fine silt. Shell material forms extremely rare layers of mostly broken shells and shell detritus with isolated speciments of large Mytilus.

The following petrographic types have been identified in recent sediments of the hydrogen sulfide zone, by A. D. Arkhangel'skiy and N. M. Strakhov: 1) deep-water gray clay; 2) transitional calcareous-argillaceous ooze (varieties α, β, and γ); 3) calcareous ooze; and 4) deep-water sand.

The distribution of these types show a certain regularity expressed in the prominence of

[1] Glinistyye mineraly osadkakh vostochnoy chasti Chernogo morya.

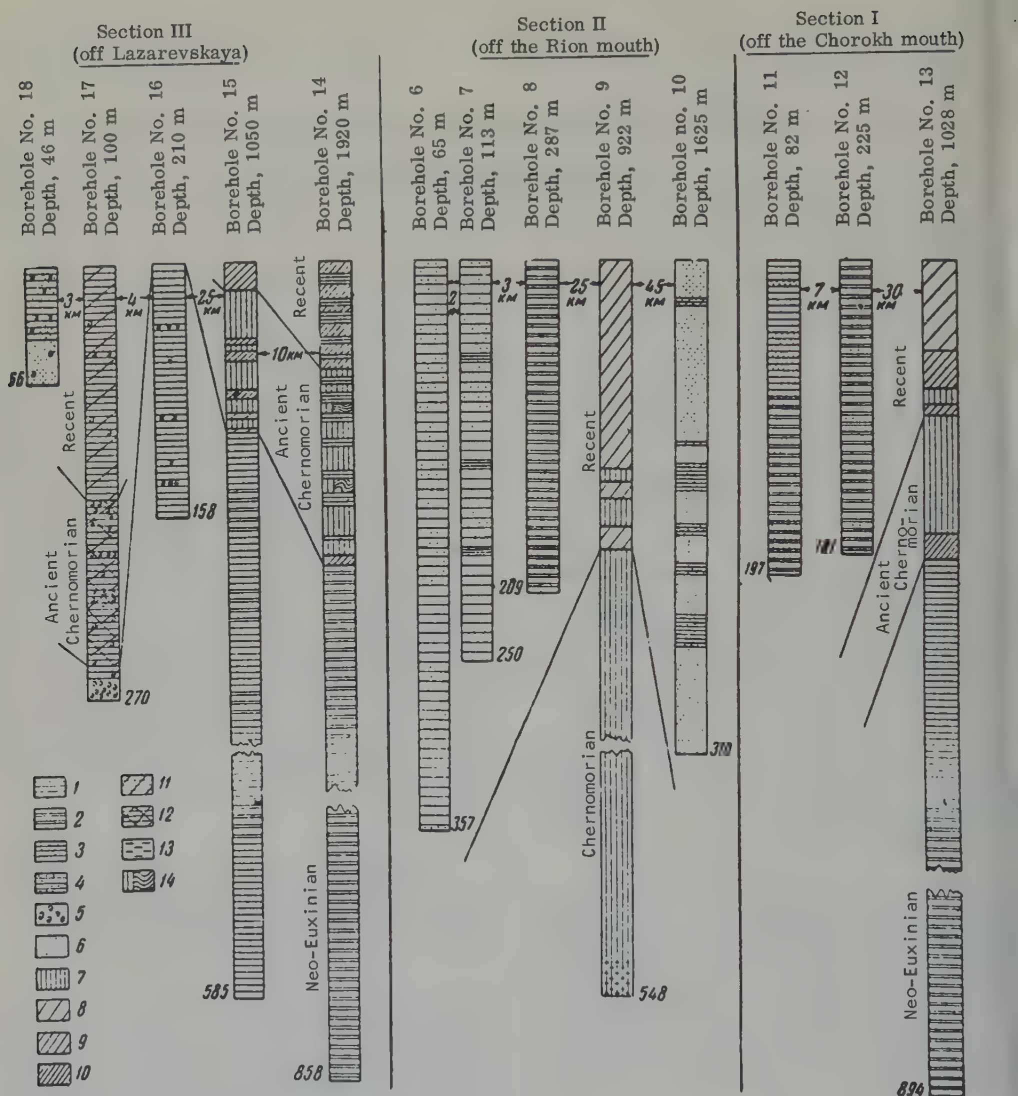

FIGURE 1. Correlation of cores by sections (core length in cm).

1 - homogeneous clay; 2 - banded to cryptostratified clay; 3 - silty clay; 4 - alternating silt and clay; 5 - shell fragments; 6 - silt; 7 - microstratified clay; 8 - transitional calcareous-argillaceous clay (variety α); 9 - same,(variety β); 10 - same (variety γ); 11 - Mytilus ooze; 12 - Ancient Mytilus ooze; 13 - isolated carbonate intercalations; Discocyclina horizons.

one or another type, depending on the character of the coast and the structure of the continental terrace. A characteristic feature of recent deep deposits developed in the area under study is the predominance of clayey varieties over calcareous, and the presence in clays of fine silt partings.

In the zone immediately adjacent to the continental terrace, at depths of 220 to 300 m, sediments consist of gray deep-water clay alone (Stations 8 and 12). This clay is marked by its high silt content, a poorly expressed stratification, and a high content of hydrotroilite. This variety has an extremely limited distribution, in our area, in sediments of the central trough developed at depths of 1000 to 2000 m. Deep-water gray clay forms here lentils from a fraction to 3-5 cm thick, which stand out from the calcareous-argillaceous ooze because of their light-gray color and homogeneous aspect.

Transitional calcareous-argillaceous ooze is xtremely widely developed in Recent deep-water ediments. This variety is very conspicuous in its tratification which was brought about by a rapid lternation of white calcareous material with a ray argillaceous and green-gray sapropelitic naterial. The thickness of individual laminae, heir sharpness, the nature of the transition, and he number per unit of measure are different for ifferent areas and horizons. Using these criteria, he transitional ooze developed in this area can be ivided into three varieties corresponding to arieties α, β, and γ of N. M. Strakhov and A.). Arkhangel'skiy.

Variety α is characterized by comparatively oor stratification but chiefly by an extremely imited development of calcareous partings which, s a rule, have no definite thickness but consist of ayered bits of drewite. Clayey and sapropelitic ayers usually do not exhibit sharp boundaries. 'ariety β is marked by a more rapid and definite tratification, with more prominent carbonate artings although morphologically they do not iffer in any way from variety α. Finally, variety has a maximum content of carbonate partings, hus approaching a pure calcareous ooze. This ariety occurs in our area only at Station 13 where here is a thin layer of it at the base of Recent de- osits.

Off the Rion mouth, at a depth of 1625 m, we dis- overed a new type of sediment, characterized by thorough development of silt with variable addi- ions of argillaceous to arenaceous material. The resence of coarse sediments in central parts of he basin is generally typical of Recent Black Sea ediments. According to A. D. Arkhangel'skiy nd N. M. Strakhov, their origin is related to seis- nicity of the Black Sea bottom, which causes strong eriodic agitation in bottom waters and sediments; s a result, coarse sand-silt particles are often arried to the central reaches of the sea.

The area under study is marked by somewhat hicker Recent sediments: 357 cm, with their ower boundary not reached by the coring; and 5 to 180 cm in the deep-water zone.

Unlike the Recent, ancient Chernomorian de- osits were found to be present only at a few points, nostly in the deep zone. In the zone of oxidation, hey were found at a single point (Station 17) of the Aytilus ooze zone. Lithologically, these deposits re in no way different from the overlying Recent nes and are represented by homogeneous gray lay with silty partings. The boundary between Recent and ancient Chernomorian layers is fixed y a sharp increase in shell material, extremely nonotonous in composition, represented chiefly y Midiae shells extremely variable in size, wall hickness, and degree of crushing. Ancient Chernomorian Mylilidae are characterized by he peculiar degree of preservation of their sually fragile valves, which is a result of their artial solution.

A bed containing such Neo-Euxinian forms as Monodacna, Micromelania, and Dressensia along with Midiae occurs at the base of ancient Cherno- morian deposits.

Only petrographic criteria are applicable in dif- ferentiating the deep-water sediments. As pointed out before, the most typical feature of Recent deep sediments is the presence of drewite intercalations which disappear in ancient Chernomorian layers; here, sediments are considerably enriched in or- ganic matter. The most common variety of ancient Chernomorian deposits in our area is micromica- ceous clay consisting of rapidly alternating micro- layers of clayey and sapropelitic material. Sandy- silty material forms very thin layers and partings. Micromicaceous clay commonly makes up ancient Chernomorian deposits and occasionally alternates with laminae of homogeneous gray clay.

White carbonate intercalations occur, as a rule, at the base of ancient Chernomorian layers; they are much less common in the middle and upper in- tervals. It is of interest that, microscopically, this carbonate material of the ancient Chernomor- ian layers is substantially different from that of Recent sediments. Instead of powder-like drewite, it is represented here by acicular, cigar-shaped to lenticular fine crystals. This peculiar habit of carbonate in ancient Chernomorian beds is an addi- tional criterion for their identification. Morpho- logically similar cigar-shaped needles of calcium carbonate were noted by N. G. Brodskaya in the Aral Sea sediments and by G. I. Bushinskiy in sedi- ments of Lake Balkhash. Both authors believe that these crystals have a strictly chemical origin.

We have observed Neo-Euxinian deposits of the shallow-water zone at a single point, in the Mytilus and ancient Mytilus ooze area, where only the upper- most and very thin portion of Neo-Euxinian sedi- ments was cored.

The boundary between ancient Chernomorian and Neo-Euxinian sediments is fairly gradual, with the change fixed by changes in biocenoses. Mylili- dae are totally missing in Neo-Euxinian deposits where they are replaced by small and extremely numerous Dreissensia shells accompanied by very widely distributed Micromelania. Off Lazarev- skaya village (Section III), Neo-Euxinian deposits are exposed at the bottom, at a depth of 210 m; they are represented by light-gray, compact, homo- geneous clay with rare intercalations of shell de- tritus and individual Dreissensia shells, intact. In deep-water cores, Neo-Euxinian deposits pre- dominate; their lower boundary has not been reached in any of the cores, although their thickness in longer cores is 4 to 6 m.

The boundary between ancient Chernomorian and Neo-Euxinian layers in the deep-water zone is quite sharp, with Neo-Euxinian layers marked by the Homogeneity of their petrographic composition and represented by light-gray clay more or less enriched in silt and sand, in different horizons. This clay is locally cryptostratified because of differences in iron hydroxide content.

Inclusions, spherules, and nodules of pyrite

are typical of Neo-Euxinian deposits. Calcium carbonate usually does not form individual layers; instead, it is diffused in the sediments. Present in the clay are isolated fragments probably of Micromelania sp., thus confirming the Neo-Euxinian age of these deposits.

Their mechanical composition was determined by ultragranulometric analysis based on the Robinson method, using sodium pyrophosphate as a disperser. All samples under study were marked by a high degree of dispersion and did not contain particles larger than 0.1 mm, with material from littoral cores being somewhat coarser than from deep-water cores. Table 1 shows that the 0.1 to 0.01 mm fraction accounts for 35 to 40% of the total in littoral samples, while it usually does not exceed 10% in deep-water samples; the intermediate fraction 0.01 to 0.001 mm makes up 30 to 40% of littoral oozes and 50 to 60% of the deep-water; and the colloid fraction proper (<0.001 mm) ranges between 35 and 50% in deep-water oozes, while its content in littoral facies is somewhat lower than 20 or 30%. It should be noted that by far most colloid particles are smaller than 0.00028 mm (0.28 micron).

Inasmuch as the bulk of clay minerals are concentrated in the <0.001 mm fraction, this fraction was studied in most detail by a combination of X-ray, thermal, optical, chemical, staining, and electron microscopic laboratory methods. A total of 30 samples of present day oozes were studied. The X-ray analysis of all samples was done in the X-ray Laboratory of the VSEGINGEO, by Ts. M. Rightburd. The absorption capacity was determined by Ye. G. Zaytseva, at the Chemical division of the Oceanography Institute. Chemical, thermal, and electron microscopic studies were carried out in corresponding laboratories of the Geological Institute, AS U.S.S.R.

In order to determine the effect of the sources of sediments on the composition and distribution of minerals in the uppermost layer of Black Sea sediments, we also studied the assemblage of heavy minerals in the silt fraction. Immersion studies have shown that the composition of the silt fraction of all samples is the same, on the whole, although there is a definite difference between the relative content of individual components in samples from the two southern sections (opposite Rion and Chorokh Rivers) and the northern section (off Lazarevskaya). In the south, predominant components of the heavy fraction are monoclinic pyroxenes represented by pale green grains marked by their fresh aspect. Minerals of the epidote-zoicite group are fairly well developed along with pyroxenes in these samples, with pure zoicite very common. Epidote grains, as a rule, show signs of partial decomposition. In addition hornblende and assorted micas with biotite predominant are fairly common in these samples. Most common among accessory minerals are zircon, tourmaline, garnet, apatite, rutile, staurolite, spinel, sphene, chlorite, and glaucophane.

The heavy mineral assemblage in the third, the northern section is similar to that of the two southern, in composition. The main difference is in its comparatively higher development of epidote which predominates in these samples, and in the composition of micas. Biotite is almost totally absent and muscovite and sericite-like micas are predominant.

On the basis of the difference in composition of clay components in the north and the south of this area of the Black Sea, this difference being in full accordance with that in the composition of heavy minerals, we tentatively designated two mineral provinces: the southern which includes the two sections opposite Rion and Chorokh Rivers and the northern, with the Lazarevskaya section. As will be shown below, the designation of these two mineral provinces is determined by a difference in the structure of their respective source areas.

The following minerals were identified by a comprehensive laboratory analysis of colloid fractions from the Black Sea oozes in this area: hydromicas, which are prominent everywhere; montmorillonite, present to some extent in almost all samples; small amounts of kaolinite, halloysite, chlorite, unaltered micas, and hydroxides of iron; present everywhere also is fine-grained calcite, opal, quartz, and biogenic silica.

The assemblage of clay minerals is fairly similar throughout the area, although there are some differences in the mineral composition of the uppermost layer from the northern and southern province. The main difference is an appreciable increase in the montmorillonite content in the south compared with the north.

Turning now to the characteristics of the clay minerals, it should be noted that Recent clay oozes are extremely difficult subjects of laboratory study because of their high dispersion, imperfect crystallization of particles, close relationship between individual mineral components, abundance of finely dispersed carbonate material, and some degree of salinity. The samples under study were processed for a fairly long time in 2% HCl, with subsequent careful washing of Cl^- and control of $AgNO_3$.

Separation of the colloid fraction was done by the standard method of elutriation in a series of beakers, with the intermediate fraction 0.01 to 0.001 mm also separated in a number of samples. A rough preliminary identification of clay minerals was done by the staining method.

Because of the long processing of the sample with HCl, the data obtained by this method are not quite reliable and yield only the most general

Table 1

Mechanical composition of sediments in the eastern part of the Black Sea

Nos.	Station Nos.	Horizon	Depth in m	Natural moisture content	Fractions in mm										
					1,0–0,5	0,5–0,25	0,25–0,10	0,10–0,05	0,05–0,01	0,01–0,005	0,005–0,001	1–0,54	0,54–0,44	0.44–0,28	0.28
1	6	17—28	65	56.46	—	—	—	30.56	29.67	8.77	12.00	3.27	1.30	0.98	13.45
2	6	120—140	65	25.98	—	—	—	42.71	21.22	6.85	13.12	2.33	2.17	0.93	10.67
3	6	120—140	65	50.99	—	—	—	16.99	35.96	9.22	15.40	5.82	1.33	1.60	13.68
4	6	262—266	65	51.76	—	—	—	8.21	43.68	10.71	16.18	3.06	3.40	2.71	12.05
5	7	173—225	113	39.95	—	—	—	0.75	32.06	13.14	26.34	4.68	5.54	0.58	16.91
6	8	Bottom Dredge	267	69.35	—	—	—	0.05	25.50	12.65	27.65	10.73	1.52	2.95	18.99
7	8	170—182	287	69.22	—	—	—	0.15	31.71	19.02	23.42	2.50	4.05	4.19	14.96
8	9	0—15	922	101.53	—	—	—	1.05	3.58	13.84	36.33	12.25	5.43	5.04	22.48
9	9	105—110	922	102.98	—	—	—	0.88	6.21	21.22	38.18	7.68	1.64	6.08	18.11
10	9	199—209	922	89.62	—	—	—	0.03	11.82	20.59	36.90	7.16	2.80	3.92	16.78
11	9	524—548	922	77.65	—	—	—	0.80	10.95	19.13	37.68	7.81	4.09	3.00	16.54
12	10	0—15	1625	103.80	—	—	—	0.05	28.07	19.94	25.89	6.93	2.63	3.11	13.38
13	10	205—292	1625	33.44	—	—	—	5.97	46.42	12.65	15.41	4.57	0.37	2.36	12.25
14	11	0—10	82	48.69	—	—	—	0.24	33.26	15.97	24.47	7.25	2.62	5.62	10.57
15	11	200—207	82	42.46	—	—	—	6.05	36.45	12.94	20.43	5.47	2.49	2.38	13.79
16	12	40—41	225	62.63	—	—	—	0.01	25.98	18.09	25.49	7.18	4.21	2.59	16.45
17	13	0—15	1028	168.63	—	—	—	0.07	2.39	15.00	39.02	11.19	3.41	3.07	25.85
18	13	100—115	1028	134.49	—	—	—	0.15	4.53	17.47	42.86	5.44	3.14	1.51	24.90
19	13	297—310	1028	54.78	—	—	—	0.01	20.98	15.73	31.40	3.68	2.78	4.25	21.17
20	13	806—816	1028	41.51	—	—	—	0.43	16.38	15.31	32.24	9.02	3.15	5.17	18.30
21	14	0—10	1920	93.25	—	—	—	0.29	14.70	12.30	33.70	9.96	6.00	2.19	20.79
22	14	70—78	1920	37.95	—	—	—	0.47	14.01	15.76	33.36	7.49	4.37	4.87	19.67
23	14	135—144	1920	69.58	—	—	—	0.48	1.47	8.16	38.08	11.88	4.10	<0.44 35.83	Coagulates
24	14	236—247	1920	63.41	—	—	—	0.12	18.95	15.19	31.51	8.50	5.24	1.95	18.54
25	14	346—356	1920	63.64	—	—	—	6.70	18.03	16.38	32.13	5.62	2.78	2.40	15.96
26	14	811—821	1920	45.31	—	—	—	15.90	5.22	17.15	25.91	7.20	4.13	5.35	19.14
27	15	Bottom Dredge	1220	162.36	—	—	—	0.55	1.44	12.10	35.30	15.12	7.23	3.75	24.51
28	15	0—15	1050	133.34	—	—	—	1.68	1.40	12.32	35.84	20.05	1.16	3.48	24.07
29	15	20—40	1050	115.18	—	—	—	1.28	10.72	8.97	36.57	7.91	7.43	2.55	24.57
30	15	388—398	1050	51.80	—	—	—	0.16	0.49	30.43	30.70	8.42	5.34	4.39	20.07
31	15	1038—1052	1050	66.00	—	—	—	2.22	7.78	10.21	34.12	8.87	6.55	3.24	27.01
32	16	0—10	210	160.56	—	—	—	0.06	12.98	17.58	31.83	9.00	5.12	3.76	19.67
33	16	146—158	210	55.06	—	—	—	0.21	25.51	12.15	28.16	7.03	4.33	2.23	20.38
34	17	2—10	100	109.02	—	—	—	0.57	12.78	16.56	28.85	10.01	2.33	5.31	23.59
35	17	211—221	100	61.91	—	—	—	8.70	2.11	16.68	37.19	7.52	5.77	1.19	20.84
36	18 (1)	55—66	46	26.82	—	—	2.35	35.99	18.95	10.27	13.96	3.97	2.10	2.53	9.88

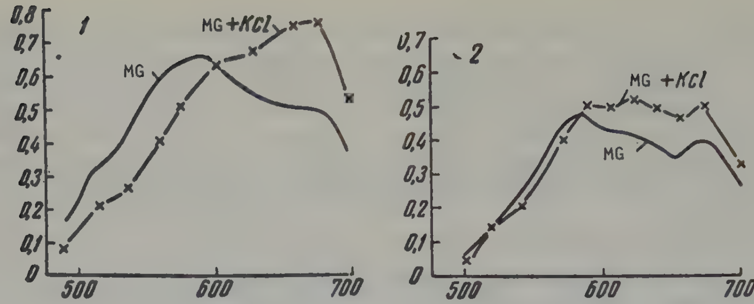

FIGURE 2. Absorbtion curves for MG (dye) and MG + KCl.

1 - sample 18, station 13, depth 800 to 810 cm; 2 - sample 1, station 14, depth 50 to 68 cm.

preliminary information on the composition of the clay minerals. A vast majority of samples so studied produced two-peak absorption curves for MG (dye) with maxima at 540 to 560°C and 660 to 670°C. With the addition of KCl, the long-wave peak grows appreciably, which suggests a hydromicaceous-montmorillonite composition for the clay components (Figure 2).

The difference in mineral composition in the northern and southern provinces stands out well in photometric curves of suspensions treated with a small amount of dye, which brings out only such strong dye absorbers as montmorillonite.

Absorption curves for samples from the southern province show a long-wave maximum obtained by the addition of KCl, considerably more intensive than that in samples from the northern province, which is a proof of the wider distribution of montmorillonite in the south (Figure 3).

Optical study of colloidal material was done on oriented aggregates of clay particles, and on unoriented crumbs of clay. In both instances the difference in optical properties between clay from the northern and southern provinces was quite definite.

Attempts to obtain well-oriented aggregates of clay particles for colloidal fractions treated with 2% HCl as a rule, were unsuccessful. It was possible to measure only the average refractive index which turned out to be considerably lower (0.010 to 0.015) than that for fractions not treated with HCl.

Aggregates of oriented particles of fractions <0.001 mm, not treated with HCl, have

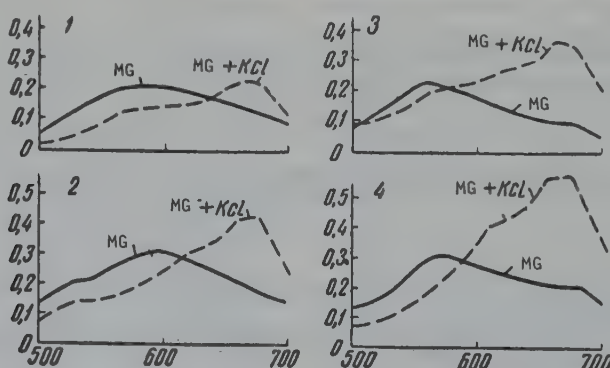

FIGURE 3. Absorbtion curves for MG and MG + KCl, with a small amount of dye.

1 - sample 1, station 14, depth 50 to 68 cm; 2 - sample 3, station 14, depth 820 to 830 cm; 3 - sample 16, station 13, depth 30 to 50 cm; 4 - sample 18, station 13, depth 800 to 810 cm.

e following refractive indices:[2] for the north-n province, γ' ranges from 1.573 to 1.561; , from 1.558 to 1.546; for the southern prov-ce, γ', from 1.555 to 1.543, dropping to 1.536 isolated instances; α', from 1.543 to 1.534, id 1.525 in isolated samples.

The samples under study have refractive in-ces somewhat lower than those for hydromicas id considerably higher than for pure mont-orillonite, which suggests a mixed hydromica-ontmorillonite composition for these fractions. ie appreciably lower refractive indices for e southern area samples graphically indicate higher amount of montmorillonite in the south-n area oozes, as compared with the northern.

A comparison of refractive indices for aggre-ites of particles of different degrees of dis-ersion (obtained after day-long and longer utriation) led to the conclusion that the bulk montmorillonite is concentrated in the finest spersed portion of the colloidal fraction.

The particle aggregates for the finest portion fraction <0.001 show refractive indices con-derably lower than those for aggregates of e entire colloidal fraction (by 0.010 to 0.015). iis result has also been corroborated by X-ray ialysis.

Studied optically, in addition to aggregates of iented clay particles, with an averaged refrac-ve index for polymineral clays, were aggre-ites of unoriented particles from the colloidal action treated with HCl, with their refractive dices ranging widely from 1.513 to 1.555.

This last circumstance strongly suggests a implex, at least a bicomponent, constitution the clay groundmass, with refractive indices ibstantially different for each component. The ghest index belongs to aggregates with the west content of low-refractive index compo-ents; and vice versa. The true refractive in-ces for each component will naturally be dif-rent from the figures cited, because the pro-irtion of the components varies from sample sample. In order to determine the relative intent of each mineral component in colloidal actions from different areas and at different ratigraphic levels, a statistical count of ains with different refractive indices was ine for four areas located in different prov-ces and at different depths.

The data so obtained were used in constructing a graph where refractive indices are plotted on the horizontal axis, and the number of aggre-gates with the corresponding refractive index on the vertical axis (Figure 4). Samples 16 and 18 were taken at Station 13, the southern area; samples 1 and 3 have higher indices (1.534 to 1.55), which shows once more [that in samples 16 and 18, montmorillonite is asso-ciated with][3] upper parts of cores; in 18 and 3, with their deeper parts.

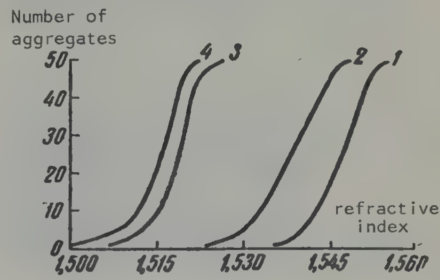

FIGURE 4. Relationship between refrac-tive indices for colloidal fractions from different provinces and different depths.

1 - sample 1, station 14, depth 50 to 68 cm; 2 - sample 3, station 14, depth 820 to 830 cm; 3 - sample 16, station 13, depth 30 to 50 cm; 4 - sample 18, station 13, depth 800 to 810 cm.

Refractive indices for samples 16 and 18 fluctuate between 1.513 and 1.525; samples 1 and 3 show higher indices (1.534 to 1.555), which confirms the earlier conclusion about the higher montmorillonite content in the southern province. The reasons for vertical changes in optical properties for the same core will be discussed below.

Thermal curves for all samples are quite similar to each other, which suggests a generally uniform composition for these sediments. All curves show three endothermal peaks: the first, always the most intensive, at 100 to 140°C; the second, at 540 to 560°C; and the third, incon-spicuous and flat, at 830 to 850°C. In addition, most curves show a small exothermal event at 910 to 930°C. The nature of these thermal curves suggests a mixed hydromica-montmoril-lonite composition for the clays.

Despite the similar aspect of these curves, they can be divided into two groups, by the

[2] It should be noted that in measuring refractive dices for aggregates of fraction <0.001 not treated ith HCl, the content of finely dispersed carbonate aterial did not affect the optical constants of the ay groundmass. This was checked in a number of imples with quite different carbonate content; their efractive indices turned out to be quite similar or early so.

[3] Translator's Note: A whole line of text is missing. His probable wording is given in brackets.

intensity of their low-temperature endothermal reaction corresponding to the loss of absorbed water.

The first group, which includes the northern area samples, has a comparatively less intensive lower endothermal peak; on curves from the southern area samples, it stands out sharper, and is connected with a higher montmorillonite content in these samples (Figure 5).

The chemical composition of the fractions studied is extremely monotonous, similar to that of typical hydromicaceous clays. Their somewhat higher content of hygroscopic water (up to 7.47%) and fairly low K_2O content, ranging from 1.34% to 2.12%, together with comparatively high SiO_2/R_2O_3 ratios (3 to 4.3), are determined by the presence of montmorillonite.

Electron-microscopic study of a large number of the Black Sea ooze samples, too, has confirmed the high uniformity of their composition (Figure 6).

The predominant mineral is hydromica represented by isometric, elongated, irregular scales of various sizes and density, and of variable sharpness in outline. Present along with clean-cut scales (a) are widely developed hydromica flakes with vague outlines (b), probably suggesting an incipient decomposition of hydromicas.

Amorphous particles of montmorillonite with vague outlines (c) are more or less conspicuous along with hydromica in all photographs.

Besides these principal minerals, hydromica and montmorillonite, the electron microscope revealed the presence of bits of kaolinite (d) and isolated elongated particles, probably of halloysite, at times displaying a well-defined tubular form (e).

In X-ray study of the Black Sea oozes, consideration was given to the extreme complexity of the subject and the apparent complete uniformity of its composition. Along with the standard method, analysis was done on samples treated with glycerine, oriented samples, and those heated to 600°C, for two hours. This comprehensive and refined analysis has brought out changes, however small, in the quantitive relationship of minerals, both areally and vertically, as well as structural differences between hydromicas present in the sediment, which in turn led to tentative conclusions on the course of the dia-genetic transformation in clay minerals from the Black Sea oozes.

X-ray photographs were obtained in chambers with diameter of 86, 114, and 57.3 mm, with copper radiation CuλK_α = 1.539 kX, and a nickel filter 0.01 mm thick.

According to X-ray data, hydromica is the predominant mineral in all samples. Line d

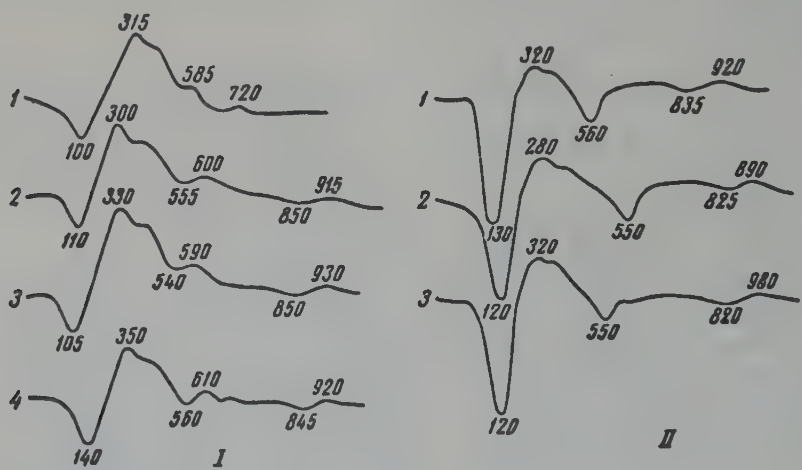

FIGURE 5. Differential thermal curves for colloidal fractions of sediments under study.

I) curves for northern area samples: 1 - sample 1, station 4, depth 50 to 68 cm; 2 - sample 2, station 14, depth 120 to 135 cm; 3 - sample 5, station 15, depth 35 to 40 cm; 4 - sample 7, station 16, depth 10 to 15 cm.

II) curves for southern area samples: 1 - sample 12, station 12, depth 158 to 165 cm; 2 - sample 18, station 13, depth 800 to 810 cm; 3 - sample 25, station 8, depth 170 to 182 cm.

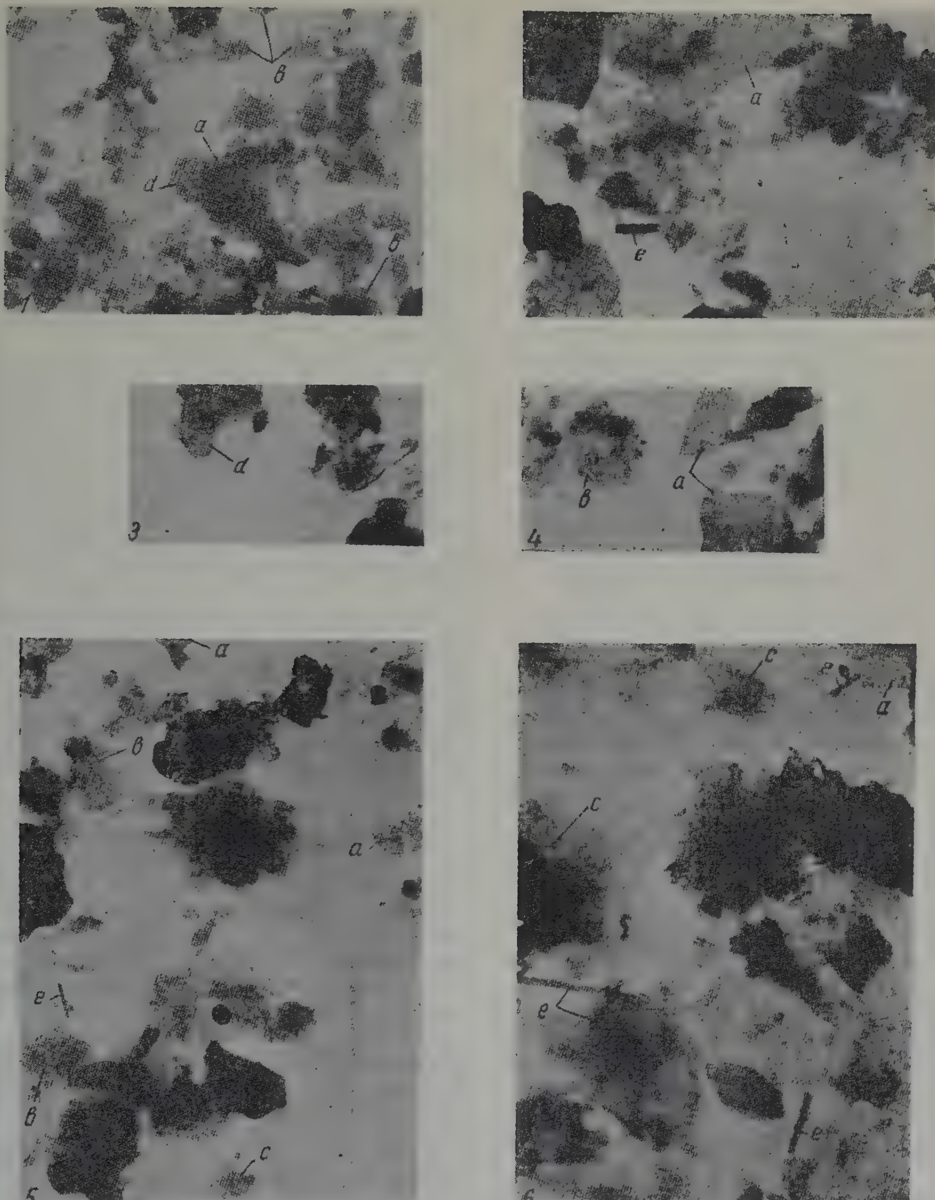

FIGURE 6. Electron-microscopic photographs of samples under study (1 - 6).

a - hydromica with sharp outlines; b - hydromica with vague outlines; c - montmorillonite; d - kaolinite; e - halloysite.

001) = 10 kX widens toward the smaller angles; t becomes somewhat sharper with heating; and t shifts somewhat when a sample is treated with glycerine. All this suggests that hydromica has a complex structure and carries some swelling ayers. X-ray photographs of oriented aggregates show a high degree of orientation in hydromicas, with a sharp boundary at about 10 kX; his apparently suggests the presence, along vith varieties carrying swelling layers, of typical well-crystallized hydromicas. Montmorillonite has been observed in almost all samples (a sharp line d (001) = 17.2 Å appears when a sample has been treated with glycerine).

A comparative estimate of the montmorillonite content in different fractions shows its uneven distribution within the colloidal fraction. The bulk of montmorillonite is concentrated in the highest dispersed portion of fraction <0.001 mm.

The montmorillonite content in the southern area samples (opposite Rion and Chorokh Rivers) is appreciably higher than in the northern area. Present in almost all samples is a kaolinite-group mineral expressed in lines d (001) = 7 Å and d (002) = 3.5 Å. In isolated samples, a sharp line 14 Å appears in heating, suggesting the presence of chlorite. All samples carry finely-dispersed quartz, with iron hydoxide present in many.

Data characterizing the capacity for ion exchange were obtained from only three samples, with one located in the northern part of the area (station 14) and two others at different depths at station 9. These exchange capacities are quite high for all three samples, and similar to each other. Nevertheless, here, too, the higher montmorillonite content in the south is expressed in somewhat higher exchange capacities which in these samples are 64.13 and 66.39 mgm equivalent to 100 gm of sample, as against a somewhat higher value for the northern sample, — 59.88 mgm equivalent to 100 gm of sample.

It follows from the above exposition that the mineral composition of present day oozes is uneven in the area under study, with regard to clay components as well as heavy minerals of the silt fraction. As pointed out before, these differences constitute the basis for differentiating this area into the north and south mineral provinces.

The reason for these differences in deposits located rather close to each other is the difference in composition of the source rocks. This relationship exists mostly because of the swift current and steep grade of Caucasian rivers which carry large volumes of clastic material directly to the sea.

The rock complex developed in the southeastern Caucasus is substantially different from that in the northwest. Jurassic and Eocene volcanic-sedimentary formations are especially well developed in the southeast, in the basin of Rion and Chorokh Rivers.

The area of bedrock drained in the northeast is rather small, being made up chiefly of Upper Cretaceous carbonate flysh and Lower Cretaceous and Upper Jurassic argillaceous flysch. Volcanic rocks are poorly developed in that area.

The abundance of volcanic material in the south promotes a comparatively large influx of montmorillonite to the sea, montmorillonite being a product of decomposition of volcanic rocks. This also is the reason for a comparatively wider distribution of pyroxenes which are abundant in extrusive and other igneous rocks, while epidote, prominent in the heavy fraction of northern cores, usually is related to crystalline and metamorphic rocks, and particularly to limestone. This is in accordance with the broad distribution of metamorphic schists and carbonate flysch in northern areas. This close connection between the composition of Recent sediments and the source rocks again emphasizes the fact that processes of mechanical removal of eroded material, as river load in suspension, are a main factor in the formation of Recent sediments, including their finest dispersion fraction.

The combination of all these data affords a fairly complete mineralogic description of the finely dispersed fraction of sediments from the eastern part of the Black Sea.

The predominant mineral of fraction <0.001 mm is hydromica which, in addition to its own elements, contains swelling elements of montmorillonite, putting them together with a group of mixed micaceous minerals quite common in clayey sections. Present in all samples also is montmorillonite whose content fluctuates appreciably, both laterally and vertically.

It is of importance that montmorillonite is concentrated, as a rule, in the finest dispersed portion of the colloidal fraction. In sediments under study, the colloidal fraction carries, besides hydromicas and montmorillonite, a sizable amount of calcite, opal, quartz, biogenic silica, and some kaolinite, halloysite, chlorite, and iron hydroxide.

These Recent sediments carry evidence of diagenetic alteration, as early as at their initial stage of formation. Inasmuch as the study of early diagenetic processes is comparatively young, their significance in the formation of clay minerals is somewhat obscure. Most students of clay minerals of Recent sediments believe that paramount in the formation of these minerals are processes of mechanical removal of material, as a river load in suspension. Whether the diagenetic processes affect clay minerals in the uppermost layer of sediments, what the nature and course of these processes are, and what their relationship to specific physiochemical conditions of the environment is, call for a profound and detailed study.

This work is an attempt to give a glimpse of vertical changes in the composition of clay minerals in a layer of sediments up to 10 m thick. Studied for this purpose material from long cores was taken from points as far from each other as possible, i.e., in different mineralogic provinces with different sources of sediments.

A comparison of samples from the deepest and the shallowest parts of cores has shown that the composition of clay components is extremely monotonous in all samples. However a slight but regular increase in the montmorillonite content with depth stands out quite definitely

against this background. Thus, detailed X-ray study of sample 16 (station 13, depth 30 to 50 cm); sample 18 (station 13, depth 800 to 810 cm); sample 1 (station 14, depth 50 to 68 cm); sample 3 (station 14, depth 820 to 830 cm) has shown a higher montmorillonite content in samples 16 and 1. In the first instance, this increase is well expressed; in the second, because of the small total amount of montmorillonite, the result is not as clear. The X-ray data have been confirmed by chemical and optical analyses as well as by ion exchange capacity and staining of the most finely dispersed portion of the colloidal fraction.

A somewhat higher amount of K_2O and a lower content of hygroscopic water has been observed in higher core samples than in the lower.

Sample 1 (station 14, depth 50 to 65 cm) contains 1.72% K_2O, 5.92% H_2O; Sample 3 (station 14, depth 800 to 830 cm): 1.42% K_2O, 6.9% H_2O; sample 16 (station 13, 30 to 50 cm): 1.42% K_2O, 6.70% H_2O; sample 18 (station 13, 800 to 810 cm): 1.83% K_2O, 7.47% H_2O.

Refractive indices for the clay fraction from surface samples are on the whole higher than those for samples from lower parts of cores. Thus the refractive index for sample 1 is 1.549 to 1.555; sample 3, 1.534 to 1.546; sample 16, 1.519 to 1.525; and sample 18, 1.513 to 1.521 (Figure 4).

Staining these samples with a small amount of dye, so as to determine the content in a fraction of strongly absorbent components such as montmorillonite, has revealed a higher content of the latter in deeper parts of cores. This also became quite apparent in a relative increase in the long-wave maximum of spectrophotometric curves, after adding KCl (Figure 3).

The cation exchange capacity also increases somewhat with depth. Thus, the sample at station 9 (depth 140 to 150 cm) has a capacity of 64.13 mgm equivalent to 100 gm while a sample from 500 cm at the same station has a capacity of 66.39 mgm equivalent to 100 gm of the fraction.

Differences between the corresponding sets of these data are so small as to be within the error of observation. Nonetheless, their combination undoubtedly suggests a tendency toward a higher montmorillonite content with depth. By the same token, the similarity of processes observed in cores distant from each other and located in areas with different sources of sediments suggests that the changes observed there are due to incipient diagenetic alteration of clay, from hydromicas to montmorillonite. The mixed stratification of hydromicaceous components, too, probably suggests an incipient "montmorillonitization" process.

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SEDIMENTARY GAPS IN THE UPPER TOURNAISIAN AND LOWER VISEAN TERRIGEOUS SECTION OF TATARIYA[1]

by

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As determined by drilling, the Volga-Ural arch [10] consists of a belt of arched uplifts separated by troughs. The Tatarian arch in question is represented by two highs in the crystalline basement, the northern (Kabyk-Kupersk) and the southern (Al'met'yevsk-Bavlinsk), separated by the Saraylinsk trough [5] which represents the northeastern part of the Kama-Kinel trough [6].

Terrigenous deposits of the Tatarian Lower Carboniferous were studied from well sections located in different tectonic environments: on limbs and axial parts of the Yelabuga trough and Birsk saddle; in the south dome of the Tatarian arch; and on the slope of this dome, toward the Melekess trough. The boreholes are lines up in two sections: I-I, crossing these tectonic elements from north to the south, and II-II running from southwest to northeast (Figures 1, 2, and 3). The lithology of terrigenous deposits, along with the spore-pollen and microfaunal assemblages from the underlying limestones, suggests that sedimentation in marginal parts of the trough and in the investigated areas of the south dome of the Tatarian arch was not always continuous. For this reason, it is wrong to regard as contemporaneous the lithologically similar members of the lower terrigenous intervals from these troughs and especially the south dome of the Tatarian arch, as has been done by many students [3].

The age of terrigenous pre-Stalinogorsk deposits from the Volga-Iral Lower Carboniferous has been different for different authors [1, 6, 7, 8]. We have adopted here the Teodorovich-Khachatryan-Sokolova differentiation of Lower Carboniferous deposits of the Middle Volga [9], according to which their lower members (Malinovsk) are Tournaisian while the upper (Radayevsk, Stannogorsk, and Lower Tula) are Visean.

In correlating the terrigenous sections of Tatariya we used L. F. Rostovtseva's identification of foraminifera and M. A. Nedoshivina's identification of microspores. We take this opportunity to express our gratitude to both of them. In addition, we used the earlier identification of spores by S. N. Naumova, L. A. Yushko, and T. V. Byvsheva.

UPPER TOURNAISIAN SUBSTAGE.

Rakovsk beds, C_1^{rak}

The Rakovsk beds, which underlie Lower Carboniferous terrigenous deposits, are the thickest in the limb exposures of the Yelabuga trough. Thus, in boreholes Yelabuga No. 19 and Bondyuga No. 48, on the north limb, these beds are represented by argillaceous-calcareous deposits, 27 to 45 m thick. Carbonate deposits at the base of the terrigenous beds have not been penetrated by boreholes in the axial part of the trough.

In Yelabuga borehole No. 19, drilled in the north limb of this trough, the Rakovsk beds are divided into three members.

The lower member (depth, 1202.1 to 1207.6 m) is represented by dark-gray to yellowish biomorphic detrital (foraminiferal) limestones with subordinate calcareous shale. The organic detritus consists of crinoids, ostracods, brachipods, trilobites, tubular algae of the Nodosinella type, and recrystallized Ungdarella.

The middle member (depth, 1198.05 to 1202.1 m) consists of dark-gray slaty shale and quartz siltstone with worm tracks. Also present are thin intercalations of argillaceous dolomitic limestone. All varieties carry sponge spicules (very common) and fragments of brachiopods (rare), ostracods, and echinoids. These rocks carry finely dispersed pyrite and a large amount of plant material. Kaolinite material is widely developed in the siltstone cement.

The upper member (depth, 1181 to 1198.05 m) is made up of dark- to light-gray nodular limestone with detritus and rare foraminifera. The nodules are round to oval but mostly irregular, 0.06 to 0.30 mm (sic) [probably m] across. They often represent foraminifera reworked by blue-green algae. The organic detritus consists mostly of tubular algae of the

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[1] K voprosu o pereryvakh v osadkonakoplenii terrigennoy topshch i verkhnego turne i nizhnego vize Tatarii.

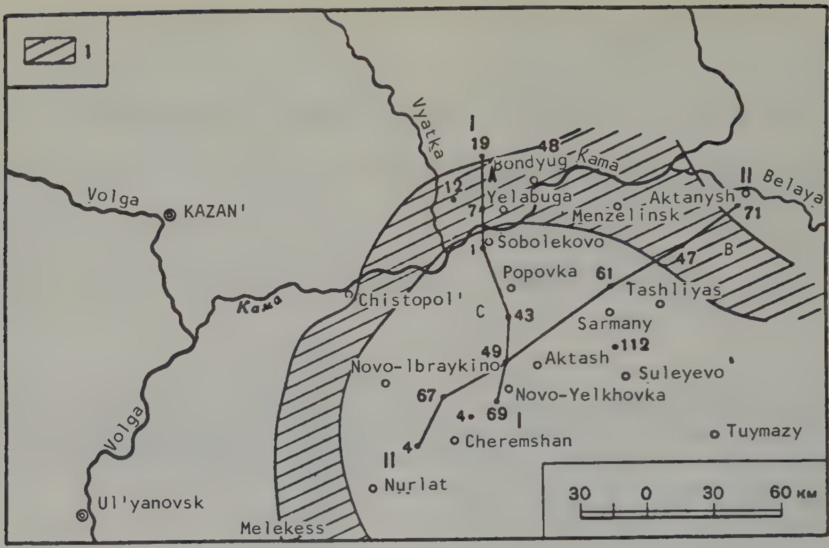

FIGURE 1. Index map of boreholes under study.

1 - zone of thick sections; A - the Yelabuga trough; B - Birsk saddle; C - southern dome of the Tatarian arch.

Nodosinella group, as well as crinoids, echinoids, brachiopods, and ostracods. Many organic remains are perforated by blue-green algae.

The following upper Kizelov foraminifera assemblage has been identified from the lower and upper members: Tournayella discoidea Dain, T. cf. Dainae Malach., T. rossica Malach., T. kisella Malach., T. regularis Malach., T. cf. gigantea Lip., T. aff. fastosa Malach., Septatournayella segmentata (Dain), S. pseudocamerata Lip., Glomospirella sygmoidalis (Raus.) Gl. irregularis (Moell.), Gl. cf. pseudopulchra Lip., Gl. spirillinoides Grozd. et Leb., Glomospira quadrata Malach., Septaglomorspiranella cf. dainae Lip., Ammobaculites cf. attenuatus Malach., Endothyra recta Lip., End. analoga Malach., End. cf. latissima Malach., End. spinosa N. Tchern., End. tenuiseptata Lip., End. kosvensis Lip., End. latispiralis Lip., End. latispiralis Lip. forma minima, End. cf. margarita Malach., End. paracostifera Lip., End. paracostifera var. multicamerata Lip., End. apta Malach., End. subrotunda Malach., Carbonella spectabilis Lip., Chernyshinella paucicamerata Lip. and a large number of Hyperaminnae. In addition, the following have been identified from the upper interval of this sequence in borehole No. 48, the Bondyuzh area: Vicinisphaera squalida Antr., Tournayella unica Malach., T. primaria Malach., Glomospira cf. formosa Malach., Endothyra cf. tuberculata Lip., End. costifera Lip., End. cf. rjausakensis Lip., End. inflata Lip., End. corona Malach., Spiroplectammina nana Lip., Chernyshinella cf. tumulosa Lip.

In the flank of the Birsk saddle, the Aktanysh borehole No. 71 penetrated higher Upper Kizelov limestone represented here by pelitomorphric siliceous limestones with detritus marked by an abundance of sponge spicules. Identified from 1317 to 1324.5 m the interval of siliceous limestone underlying the terrigenous deposits were Tournayella discoidea, Dain forma maxima T. discoidea Dain, T. kisella Malach., T. subangulata Malach., T. primaria Malach., Endothyra paracostifera Lip., End. tenuiseptata Lip. Quasiendothyra urbana Malach., Glomospiranella irregularis (Moell.). These limestones change upward to Malinovsk terrigenous beds with intercalations of siliceous spicule limestones in their lower part.

The absence of siliceous limestones with sponge spicules in the northern limb of the Yelabuga trough, in borehole No. 19, and the sharp contact with terrigenous deposits, suggest that the Rakovsk beds are not fully represented here but are separated by a break from the overlying terrigenous beds, upper Malinovsk, judging from their spore content.

Although limestone underlying the terrigenous sequence has not been penetrated in the axial part of the Kama-Kinel trough (Yelabuga borehole No. 4) and Birsk saddle (Menzelinsk-Aktanysk borehole No. 47), beds of argillaceous limestone with sponge spicules were found in the lower part of the Malinovsk beds.

In the area of shorter intervals, in the south dome of the Tatarian arch, the terrigenous sequence also rests on upper Kizelov limestone whose thickness has been reduced here to

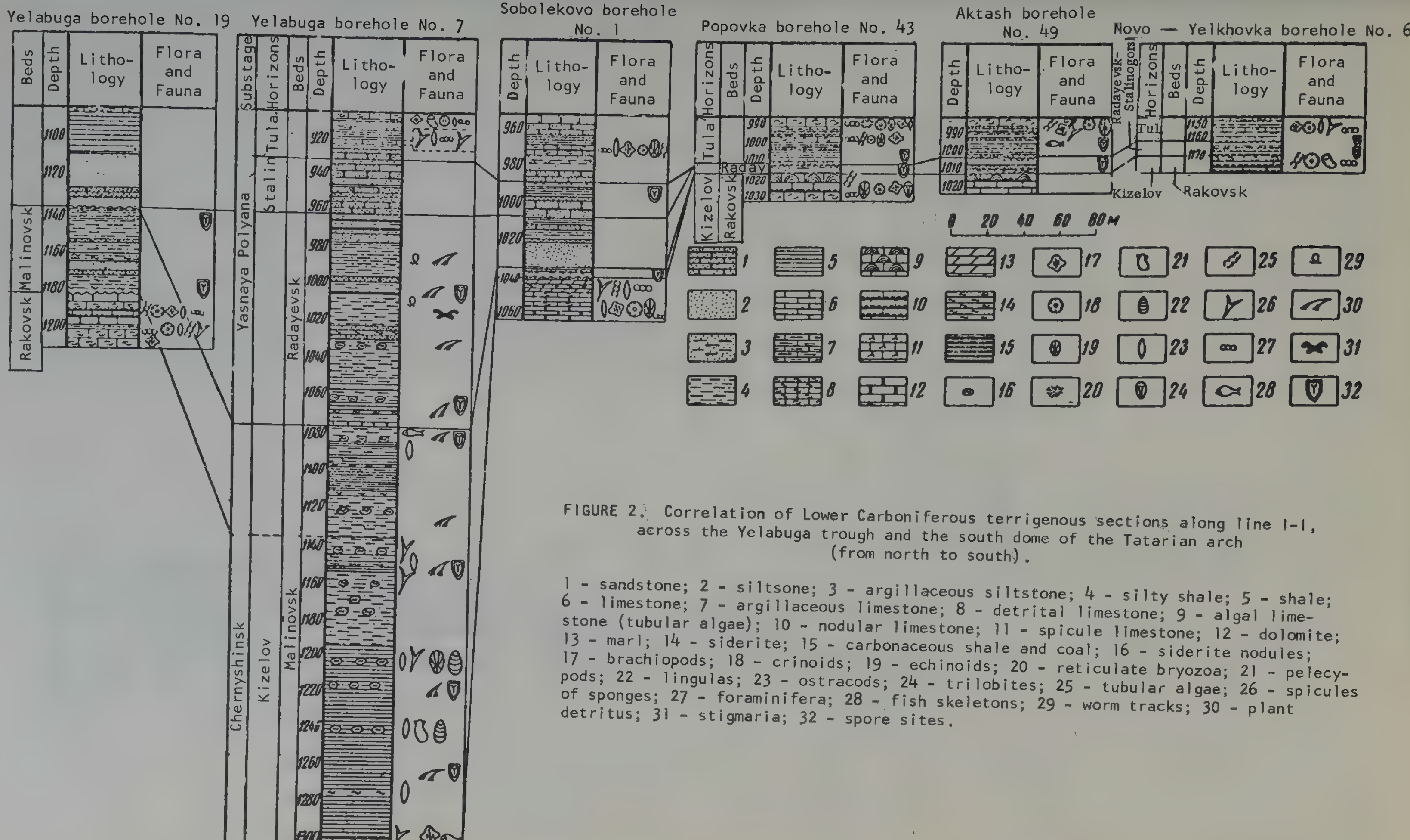

FIGURE 2. Correlation of Lower Carboniferous terrigenous sections along line I-I, across the Yelabuga trough and the south dome of the Tatarian arch (from north to south).

1 - sandstone; 2 - siltsone; 3 - argillaceous siltstone; 4 - silty shale; 5 - shale; 6 - limestone; 7 - argillaceous limestone; 8 - detrital limestone; 9 - algal limestone (tubular algae); 10 - nodular limestone; 11 - spicule limestone; 12 - dolomite; 13 - marl; 14 - siderite; 15 - carbonaceous shale and coal; 16 - siderite nodules; 17 - brachiopods; 18 - crinoids; 19 - echinoids; 20 - reticulate bryozoa; 21 - pelecypods; 22 - lingulas; 23 - ostracods; 24 - trilobites; 25 - tubular algae; 26 - spicules of sponges; 27 - foraminifera; 28 - fish skeletons; 29 - worm tracks; 30 - plant detritus; 31 - stigmaria; 32 - spore sites.

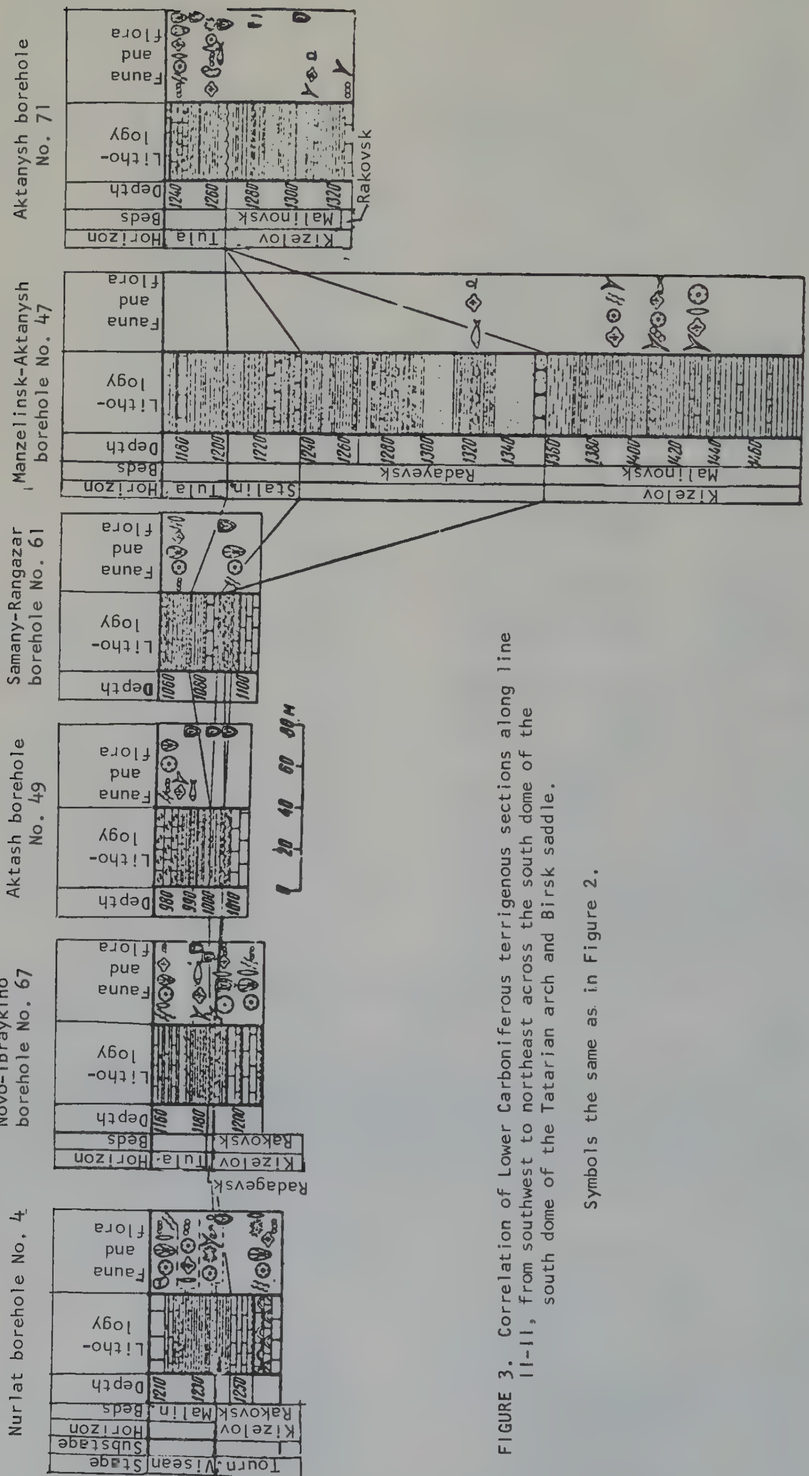

FIGURE 3. Correlation of Lower Carboniferous terrigenous sections along line II-II, from southwest to northeast across the south dome of the south dome of the Tatarian arch and Birsk saddle.

Symbols the same as in Figure 2.

- 12 m. This limestone is gray to light-gray nd brown, cavernous, nodular, detrital, and aturated with oil. Nodules of blue-green algae, 06 to 0. 36 m, are round, oval, to irregular. lany nodules are made up fully of pelitomor-nric carbonate, while the others carry re-orked detritus with foraminifers. The nodules re in close contact with one another, with the pace between them filled up by secondary re-rystallized calcite with bituminous inclusions. he organic remains are of crinoids, echinoids, stracods, brachiopods, corals, and tubural lgae of the Nodosinella group. Many of them re pierced by blue-green algae, with all stages f their reworking by the latter visible. Nodular mestones are commonly underlain and overlain y mixed algal limestone consisting of the tubu-ır algal remains (Nodosinella ?) as well as odules, or of the remains alone.

Nodular limestones are widely developed over ie south dome of the Tatarian arch, in the fol-owing areas: Novo-Ibraykino, Novo-Yelkhovka, urlat, Aktash, Popovka, and Sarmanovo-angazar. The following foraminifera were iden-fied from mixed algal limestone (depth, 1156 to 168. 9 m) in borehole No. 69, the Novo-Yelk-ovka area: Tournayella discoidea Dain, T. Dis-oidea var. angusta Lip., Septatournayella? iinuta Lip., Glomospiranella irregularis Moell), Gl. pseudopulchra Lip., Glomospira pirillinoides Grozd. et Lib., Endothyra ex gr. iberculata Lip., End. kosvensis Lip., End. cf. ostifera Lip., End. cf. recta Lip., End. cf. enuiseptata Lip., Quasiendothyra cf. urbana lalach., Chernyshinella glomiformis Lip. forma iinima, Ch. paucicamerata Lip., Spiroplectam-iina parva N. Tchern. In addition, Ammodisucs ellus Malach., Quasiendothyra cf. urbana lalach., Endothyra recta Lip., and E. costifera ip. were determined from borehole No. 112, ie Suleyevo-Tashlyar area (depth, 1175. 3 to 82. 3 m). According to L. F. Rostovtseva, this oraminiferal assemblage suggests an upper izelov age for the enclosing limestones.

Thus, limstones at the base of the terri-enous sequence, in all boreholes, are upper izelov rather than lower Kizelov as believed y M. L. Kiligina and A. K. Shel'nova [3].

Upper Kizelov nodular and tubular-algal mestones of the Tatarian arch are most likely orrelative with the Rakovsk interval from the mb of the trough. This is corroborated by ie lithologic similarity between these lime-tones and the upper member of the Rakovsk nterval from borehole No. 19, the Yelabuga rea, as well as by the similarity in their iicrofaunas. The relatively poor microfaunal ssemblage from the Tatarian arch sections is xplained by the fact that here the bulk of oraminifearal tests have been intensively re-orked by algae, and unidentifiable for that eason.

Nodular and mixed algal limestone directly underlie terrigenous deposits in most of the wells of the Tatarian arch south dome. An exception is the section in borehole No. 67, the Novo-Ibraykino area, where nodular limestones are overlain by spicule limestones changing up-ward to shale carrying sponge spicules. As a rule, however, the spicule limestone member drops out of the section, as it does on the flank of the Yelabuga trough. The absence of this member in most boreholes on the south dome, as well as in lower beds of the terrigenous sequence, suggests a sedimentary break at the upper Kizelov-Visean boundary (the absence of the Malinovsk beds and a direct contact between the Radayevsk and Rakovsk beds was recognized in borehole No. 43, the Popovka area; No. 49, the Aktash area; No. 69, the Novo-Yelkhovka area, etc.).

The Malinovsk beds — C_1^1 malin

The presence of the Malinovsk beds has been determined from spores, in the Yelabuga trough zone (borehole No. 7) and the Birsk saddle (the Aktanysh borehole No. 71). On the north flank of this trough (Yelabuga borehole No. 19), these beds are represented by their upper interval. The Malinovsk beds have been penetrated in some boreholes of the Tatarian arch south dome (Nurlat borehole No. 4; Cheremshan borehole No. 4; Novo-Ibraykino borehole No. 67).

In the axial part of the Yelabuga trough borehole No. 4) these beds are 200 to 250 m thick. They are rpresented by thin black shale with nodules of rust-brown siderite. This shale is micromicaceous, with splothes of organic matter and with finely dispersed pyrite and siderite. It also carries fine plant detritus and micro- and macro-spores. The terrigenous content (quartz and mica) increases upward, along with the appearance of numerous silty intercalations and thin layers of carbonaceous shale and coal. At the base, the shale is calcareous and carries remains of ostracods, pelecypods, lingulas, and sponge spicules. The highest concentration of ostracods and calcitized sponge spicules occurs in sideritic layers and nodules, whose number decreases upward. Siderite is associated here with marine shales.

In the axial part of the Birsk saddle (the Menzelinsk-Aktanysh borehole No. 47) the Malinovsk shale carries numerous layers of argillaceous limestone with organic detritus (sponge spicules, tubular algae of the Nodosinella group, crinoids, ostracods, and brachiopods).

In the flank area of the Birsk saddle (Aktanysh borehole No. 71), the Malinovsk beds thin down to 56 m. They consist of dark-gray calcareous shale interbedded with spicule limestone in their lower part and with siltstone and carbonaceous shale in the upper. Sideritic layers are missing and sideritic nodules are rare. The shales have

hydromicaceous, siliceous, with kaolinite beds. They contain much carbonized and gelled plant detritus and micro- and macro-spores, with abundant Stigmaria in the vicinity of coal layers. Sponge spiculas, ostracods, and fragments of brachiopods were observed in the lower interval. These faunal remains and limestone intercalations suggest a close relationship between the Malinovsk beds and Rakovsk limestone.

In the north flank of the Yelabuga trough (borehole No. 19), an upper Malinovsk interval rests directly on the Rakovsk limestone. It consists of dark-gray sideritic shale with subordinate quartz siltstone, and with carbonaceous shale toward the top.

On the south dome of the Tatarian arch (the Novo-Ibraykino borehole No. 67), the Malinovsk beds are probably represented by a three-meter thick member of black shale with sponge spicules and abundant pyrite nodules. The amount of spores obtained from this shale was insufficient for an age determination.

The remaining sections of the Malinovsk beds we studied contained a spore assemblage typical of the so-called lower Malinovsk interval of V. M. Pozner.

Predominant in the Yelabuga No. 7 borehole, at depth 1075-1223.4 m are small spores, of group Leiotritetes (16%), without a fringe and with a smooth exine, together with a spore of the Acanthrotriletes group (18%), having a thorny exine and the warty exine Lophotriletes (8%). Also present are spores with a wide fringe, the Euryzonotriletes group (11%); a filmy fringe, the Hymenozonotriletes group (12%); a narrow fringe, the Stenozonotrilites group (14.5%); and spores with a trilobate fringe, of the Trilobozonotriletes group (10%). Almost totally absent are spores with dotted ornamentations of the fringe and with a reticulate exine. Always present are a few spores with a fringe and a warty exine. The following spore species predominate: Leiotriletes subintortus (Waltz) Naum. (up to 11%); Acanthotriletes tenuisetosus Juschko (in litt.) (up to 11%); Lophotriletes inflatus Naum. (up to 11%); Euryzonotriletes subcrenatus (Waltz) Naum. (in places up to 12.5%); Hymenozonotriletes ex gr. genuinus Juschko (in litt.) (in places up to 25%); Stenozonotriletes minimus Juschko (in litt.) (up to 10.5%); St. stenozonalis (Waltz) Naum. (up to 3.5%); Archaezonotriletes decipiens Juschko (in litt.) (up to 3.5%); and Trilobozonotriletes insciso-trilobis (Waltz) Naum. (up to 1.5%).

Identified from the 1287.2 to 1295.0 m interval of the Aktanysh borehole No. 71 was another spore assemblage typical of the Malinovsk beds. It is characterized by development of fringeless spores with a smooth exine: Leitotriletes platyrugosus (Waltz) Naum., (up to 26%); L. inermis (Waltz) Naum. and L. laevis Naum. (together, up to 9%); small spores with a thorny exine: Acanthotriletes erinaceus (Waltz) Naum. (up to 25%); A. parvispinosus (Luber) Naum., A. restispinosus (Luber) Naum., and A. spinosus Naum. (together up to 6.5%); spores with a warty exine: Lophotriletes macropunctatus Naum. and L. scrobiculatus Naum. (togehter, up to 6.5%), L. minor (Waltz) Naum. (up to 6.5%). Spores with a wide fringe account for only a small percentage of the total. Most consistent among them are Euryzonotriletes incisus sp. n. (up to 1%), and E. subcrenatus (Waltz) Naum. (up to 0.5%). Spores with a filmy fringe include Hymenozonotriletes auranticus Naum. comb. n. (1%) and H. altus sp. n. Spores with a dotted fringe and a narrow fringe are rather uncommon. Characteristic among concave and trilobate spores are Simozonotriletes stenomarginatus Naum. (up to 2.5%) and Trilobozonotriletes inciso-trilobus Naum. (up to 4%).

A similar spore-pollen assemblage has been observed in borehole No. 19, Yelabuga area. However, it differs somewhat from the spores listed above by the appearance of a small number of spores of Hymenozonotriletes pusillus (Lbr.) (up to 4%) and Trematozonotriletes punctatus Naum. (up to 5%), typical of younger, i.e., the Stalinogorsk, deposits.

From carbonaceous shale directly overlying the upper Kizelov limestone in borehole Nurlat No. 4, on the slope of the Tatarian arch dipping toward the Melekess trough, T. V. Byvsheva identified a Malinovsk spore assemblage corresponding to the second (II) spore-pollen assemblage of the Melekess borehole No. 1 [2]. The predominant spores are Leiotriletes inermis Naum., Lophotriletes rugosus Naum., Acanthotriletes erinaceus Naum., with a few Trilobozonotriletes inciso-trilobus Naum.

A similar spore-pollen assemblage was determined by S. N. Naumova from the Cheremshan No. 4 borehole: Leiotriletes inermis (Waltz) Naum., L. subintortus (Waltz) Naum., Lophotriletes rugosus Naum., Archaezonotriletes reticularis Naum., Ar. tuberculatus Naum., Trilobozonotriletes inciso-trilobus Naum., T. simplex Naum.

The composition of spore assemblages from these boreholes suggests that the Malinovsk beds are correlatives of the Kizelov unit. These spore assemblages can be correlated with the second (II) and third (III) spore assemblages established by T. V. Byvsheva for the Melekess control test, and with A. M. Loginova's first and second spore assemblages from the Stalingrad Volga region [4], corresponding to V. M. Pozner's Lower Cretaceous sequence.

The Malinovsk beds are missing on the Tatarian arch south dome, in boreholes Novo-Yelkhovka No. 69, Aktash No. 49, and Popovka No. 43, where shale with a Radayevsk spore

semblage rests directly on upper Kizelov algal nestone.

Thus the Malinovsk beds carrying a Kizelov ore assemblage are not developed everywhere. his member is missing in the most uplifted rtion of the Tatarian arch south dome.

THE VISEAN STAGE

Radayevsk beds — $C_1^{2\ rad}$

We have determined the presence of the adayevsk beds by spores from boreholes Yelaiga No. 7 and Murzikha No. 12, in the Yelabuga ough; and from the Novo-Ibraykino No. 67, povka No. 43, and Aktash No. 49, in the south me of the Tatarian arch.

The Radayevsk beds have a rhythmic arrangeent in the Yelabuga trough zone, being reprented by four rhythms, 10 to 32 m thick. Lying the base of the two lower rhythms are gray iartzose, calcareous, sandy siltstone, with ieven-grained sandstone at the bases of the two oper rhythms. Toward the top, these sandylty rocks change to argillaceous siltstone inrbedded with hydromicaceous shale. The iale carries finely-dispersed pyrite and sidere, also thin sideritic lenses and intercalations. also contains much plant detritus and microid macro spores. Layers of carbonaceous iale and coal occur in upper parts of the iythms. The Radayevsk beds are variable in ickness (50 to 100 m). Unlike the Malinovsk eds, they are unfossiliferous, with the exeption of worm-tracks. While terrigenous iartz material appears in upper intervals of e Malinovsk beds, as subordinate thin interilations, the sand and silt content increases iarply because of intensified tectonic moveents. In addition, they make up a layer or ven individual members at the base of the iythmic sequence. Regional uplifts in the eposition areas and on individual arches, in the ournaisian-Visean transition time, increased e amount of sandy-silty deposits in Radayevsk me as compared with the Malinovsk.

Identified from the 988.4 to 1048.2 m interval the Radayevsk beds in the Yelabuga No. 7 orehole was a spore assemblage differing from e Malinovsk in having a lower content of fringess spores with a smooth exine, e.g., Leiotriles (9%), and spores with a filmy fringe, e.g., ymenoxonotriletes (3%), along with an increase the amount of fringeless spores with a leatherke exine, e.g., Trachytriletes (10.5%) and orny exine, Acanthotriletes (29%). Spores th a fringe and a warty exine disappear cometely.

A high content of spores Acanthotriletes uisetosus Juschko (in litt) (up to 30%), A. spinosus Naum. (up to 10%), Trachitriletes punctulatus (Waltz) Naum. (up to 7%); T. lasius (Waltz) Naum. (up to 9.5%), Euryzonotriletes incisus Jaschko (in litt.) (up to 3.5%), Stenozonotriletes laevigatus Naum. v. giganteus v. n. (up to 6%), S. laevis Naum. (up to 6.5%) is characteristic. There is a decrease in the content of such spores as Hymenozonotriletes ex gr. genuinus Jusckho (in litt.), Lophozonotriletes malevkensis Naum., L. rarituberculatus (Luber) Naum., Archaezonotriletes decipiens Juschko (in litt.), occurring in the Malinovsk beds.

A similar Radayevsk section is present in the axial part of the Birsk saddle (Menzelinsk-Aktanysh borehole No. 47). It is possible that the Radayevsk beds, and unquestionably the Stalinogorsk beds, are missing in the Aktanysh No. 71 section, on the flank of the saddle, where the Malinovsk beds appear to be overlain directly by lower Tula deposits.

The boreholes drilled in the zone of reduced thicknesses of the terrigenous sequence, on the Tatarian arch south dome, the Radayevsk beds are represented by thin (2 to 6.5 m) dark-gray to almost black silty shale, commonly saturated with organic matter, pyritic, material and carrying abundant plant detritus and spores. Subordinate quartzose silt, with thin coals in the upper interval, is also present.

Identified in the Radayevsk beds (1008.1 to 1014.5 m depth) from borehole No. 43, the Popovka area, was a spore assemblage characterized by fringeless forms with a thorny exine: Acanthotriletes erinaceous (Waltz) Naum. (17.5%), A. punctatus Juschko (3.5%), A. Parvispinosus (Luber.) comb. n. (8%), fringeless spores with a smooth exine: Leiotriletes subintortus (Waltz) Naum. (8.5%); spores with wide fringe Euryzonotriletes subcrenatus (Waltz) Naum. (11%); narrow Stenozonotriletes amplectrus Naum. (5%); spores witha filmy fringe: Hymenozonotriletes ex. gr. auranthicus Naum. comb. n. (7%); and spores with a dotted fringe: Trematozonotriletes stenomarginatus Juschko (12%). A similar sporepollen assemblage was identified from the Aktash No. 49. In these boreholes, the Radayevsk beds rest directly on the Rakovsk limestone.

Identified by T. V. Byvskeva from dark-gray shale cored at the 1184.2 to 1186.6 m in the Novo-Ibraykino borehole No. 67 was a Radayevsk spore assemblage, with Euryzonotriletes subczenatus Naum., Euryzonotriletes macrodiscus (Waltz) Naum., E. megalothelis Waltz, Hymenozonotriletes genuinus Jusch. (in litt), Trilobozonotriletes inciso-trilobus Naum., Trematozonotriletes gibberosus var. polyzonalis Jusch. (in litt.) predominating, and with a small amount of Archaezonotrilites macrotuberculatus sp. n. and Hymenozonotriletes giganteus sp. n. The spore assemblage from this shale section is correlative with the fourth (IV) assemblage from

the Melekess control test, in turn corresponding to the main upper Malinovsk interval of V. M. Pozner.

Resting on the Rakovsk upper Kizelev limestone in borehole No. 69, the Novo-Yelkhovka area is a (6 m thick) black thin-bedded pyritic shale unit with a Radayevsk-Stalinogorsk spore assemblage in its upper part; this unit belongs to the lower sub-horizon (sic) of the Stalinogorsk horizon (sic). As in borehole 43 of the Popovka area, the Malinovsk beds are missing in borehole No. 69.

Thus the shale member ("a" of M. A. Kiligina and A. K. Shel'nova [3]), which rests directly on the Rakovsk limestone in sections of the Tatarian arch south dome, is not of the same age everywhere. In some boreholes (Nurlat No. 4, Cheremshan No. 4), it carries a Malinovsk spore assemblage; in others, the Radayevsk (the Popovka No. 43) and possibly the Radayevsk-Stalinogorsk (Novo-Yelkhovka No. 69).[1] It is possible that the magnitude of the hiatus between the Rakovsk limestone and overlying terrigenous deposits increases from the periphery to the crest of the Tatarian arch south dome.

STALINOGORSK HORIZON — $C_1^{2\ st}$

Stalinogorsk deposits in the Yelabuga trough, studied from sections in boreholes Yelabuga No. 7, Murzikha No. 12, and others, have a rhythmic sequence, with two large rhythms, 15 and 20 m thick, identified in the No. 12. The base of the rhythms is represented either by gray calcareous siltstone (lower rhythm) or sandstone (upper rhythm), which change upward to dark-gray silty to carbonaceous shale with subordinate siltstone and spore-bearing coals. Kaolinite shale has often been observed in the vicinity of these coal beds. All rocks in the upper parts of a rhythm contain much organic remains and micro- and macro-spores. Unlike the Radayevsk beds, siderite in nodules, layers, and fine dispersion is altogether missing here. The upper carbonaceous shale interval is best developed in the lower rhythm. The upper rhythm, on the other hand, is characterized by the development of the lower sandy interval and a poor development of the upper, argillaceous, interval.

Stalinogorsk deposits are missing from the section of the south dome of the Tatarian arch, in boreholes Novo-Ibraykino No. 67, Popovka No. 43 and Aktash No. 49. In these boreholes Tula terrigenous deposits lie directly on the Radayevsk beds. Identified from the 1152.5 to 1160 m interval of the No. 69, the Novo-Yelkhovka area, was a Radayevsk-Stalinogorsk spore assemblage typical of the lower Stalinogorsk sub-horizon. Typical of this assemblage is the considerable increase in spores with dotted fringes, from the Trematozonotriletes group, also in spores of the Stenozonotriletes group with a narrow fringe as well as spores with a filmy fringe, Hymenozonotriletes pusillus (Lbr.) Waltz. Fringed spores with a thorny (Acanthotriltes) to warty (Lophozonotriletes) exine are less numerous than in the Radayevsk spore-pollen assemblage.

Deposits of the Stalinogorsk horizon proper are missing in borehole No. 69, the Novo-Yelkhovka area, with lower Tula deposits resting directly on its lower sub-horizon.

The absence of Stalinogorsk deposits proper, as well as of the underlying Radeyevsk beds, has been determined for the Aktanysh borehole No. 71. Here, the Malinovsk beds appear to be overlain by lower Tula deposits characterized by a Tula assemblage of spores and microfauna. F. M. Pozner assumes that Stalinogorsk deposits are missing in boreholes Buldyrsk No. 4 and Aktash No. 22 [6].

The examples cited demonstrate the absence of Stalinogorsk deposits in the most uplifted part of the Tatarian arch south dome as well as on flanks of the Birsk saddle. The absence of the Stalinogorsk unit proper, in these areas, is related most likely to a break in sedimentation rather than to erosion, inasmuch as no trace of erosion has been observed.

LOWER TULA DEPOSITS — $C_1^{2\ tul.}$

The lower interval of the Tula unit is made up, as a rule, of terrigenous formations represented by two rhythms. Tula deposits usually begin with a bed of gray calcareous sandstone, less commonly siltstone. Unlike Stalinogorsk rhythms, the Tuls rhythms are complete, consisting of both the transgressive and regressive cycles, with the first one culminating in a fossiliferous limestone. The Tula fauna is represented by fragments of brachipods, ostracods, sponge spicules, fish remains, less common crinoid segments, shreds of reticulate bryozoa, solitary corals, and occasionally by foraminifera. Brachipod fragments are common in sandy-silty rocks at the base of the rhythm.

The following foraminifera were identified in lower Tula deposits from the 1258 to 1260 m depth in borehole No. 71, the Aktanysh area: Glomospirella cf. irregularis (Moell.), Archaeodiscus krestovnikovi Raus., A. krestovnikovi var. pusilla Raus., Endothyra cf. prisca Raus. et Reitl., with the following spore-pollen assemblage identified from the 1250 to 1258 m depth: Hymenozonotriletes pusillus (Lbr.) Naum. + H. altus Juschko (35%),

[1]Spores were identified from 3 m, above the base of the shale; the base of this member may be correlative with the Radayevsk beds.

Trematozonotriletes punctatus Naum. (4.5%), T. bialatus (Waltz) Naum. (10%), T. vaelculosus (Waltz) Naum. (2%), Euryzonotriletes subcrenatus (Waltz) Naum. (5%), E. macroduplicatus Naum. (1.5%), Simozonotriletes trivialis Naum. (4.5%). Stenozonotriletes literatus (Waltz) Naum. (2.5%), Archaezonotriletes spinosus Juschko (2.5%), Trilobozonotriletes inciso-trilobus (Waltz) Naum. (3.5%), Leiotriletes graber Naum. (4.0%), L. platyrugosus (Waltz) Naum. (4.5%). L. laevis Naum., L. inermis (Waltz) Naum. (2%), Trachitriletes punctulatus Naum.) (1.5%).

In the Tatarian arch south dome, in borehole No. 67, the Novo-Ibraykino area, lower Tula deposits, too, are rhythmic. The section consists of two rhythms: the lower, 3.6 m thick; and the upper, 4 m thick. They both begin with dark-gray sandy siltstone with an argillaceous-calcareous cement, bituminous splotches, plant detritus, and macro- and micro-spores. Above them there are dark-gray argillaceous siltstones with abundant plant detritus and occasional leptochloritic oölites and calcareous radiolites. Both varieties carry many worm tracks. Occurring in the upper part of the rhythms are black montmorillonitic and hydromicaceous-montmorillonitic shales with pyritized plant and fish remains. The lower rhythm terminates with gray microgranular limestone carrying calcitized spicule remains of siliceous sponges and brachiopod fragments. Black shale of the upper rhythm contains a layer of greenish-gray marl full of plant detritus and brachiopod (?) fragments.

A Tula spore assemblage has been identified in siltstone of the lower and upper rhythms, at a depth of 1172.9 to 1186.4 m. Along with a higher content of fringeless spores, this assemblage is characterized by spores with a filmy fringe, Hymenozonotriletis pusillus (Lbr.) Naum. and spores with dotted ornamentations from the Trematozonotriletes group. Prominent are the following: Hymenozonotriletes pusillus (Lbr.) Naum. + H altus Juschko (37.0%), Trematozonotriletes bialatus (Waltz) Naum. (9%), Tr. valeculosus (Waltz) Naum. (2.5%), Trilobozonotriletes inciso-trilobus (Waltz) Naum. (4.5%), Leiotriletes graber Naum. (8%), Acanthotriletes erinaceus (Waltz) Naum. + A. tenuisetosus Juschko (1.5%), Euryzonotriletes macroduplicatus Naum. (2%), E. subcrenatus (Waltz) Naum. (2%), Stenozonotriletes laevigatus Naum. var. giganteus v. n. (3%), S. laevis Naum. (2%).

A similar spore-pollen assemblage was identified in borehole No. 3, Popovka (depth 1005 to 1014.5 m) and the No. 49, the Akhtash area (991.8 to 998 m depth).

Lower Tula terrigenous deposits are the most consistent among Lower Carboniferous clastics: they occur in all boreholes of the Yelabuga trough and Birsk saddle as well as on the Tatarian arch south dome. However, their thickness decreases from 20 to 22 m in the trough and Birsk saddle to 6 to 8 m on the Tatarian arch.

In summing up, the following conclusions can be drawn:

1. The Rakovsk beds on the flanks of the Yelabuga trough and on the Tatarian arch south dome are upper Kizelov. Their sharp contact with the overlying terrigenous deposits in the same areas is explained by the absence of the upper part of the Rakovsk beds (spicule limestone) and the lower part or all of the Malinovsk beds. In sections from the Birsk saddle boreholes and some boreholes on the Tatarian arch south dome, where both groups of beds are present, a gradual change from the Rakovsk limestones to Malinovsk argillaceous rocks has been observed.

2. The absence of the upper Rakovsk beds and some or all of the Malinovsk beds suggests a sedimentary break in the second half of the upper Kizelov, at its boundary with the Visean, in the most uplifted portion of the south dome and in the northern flank of the Yelabuga trough.

3. Going from the periphery to the most uplifted parts of the Tatarian arch south dome, the Rakovsk limestone is overlain by progressively younger deposits with different spore-pollen assemblages. It follows that the hiatus between the Rakovsk limestone and overlying terrigenous deposits grows larger toward the crest of the dome, embracing an interval from the Malinovsk through the Radayevsk beds.

4. The rhythmic constitution of the Radayevsk, Stalinogorsk, and Lower Tula deposits suggests the increasing frequency and intensity of tectonic oscillatory movements at the beginning of the Visean. This led to a sedimentary break in the most uplifted part of the Tatarian arch south dome, in Stalinogorsk time, and on the Birsk saddle slope, in Radayevsk-Stalinogorsk time.

5. Thus the deposition of the Lower Carboniferous terrigenous sequence was not continuous on the flanks of the Yelabuga depression and Birsk saddle, nor on the higher parts of the Tatarian arch south dome. There were at least two sedimentary breaks: 1) in the second half of upper Kizelov time (approximately at the Tournaisian-Visean boundary); and 2) a local one, during Stalinogorsk time.

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NEOGENE DEPOSITES OF CENTRAL KYZYL-KUM[1]

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The stratigraphy of Neogene continental deposits in Kyzyl-Kum is a subject of considerable controversy. The findings of Apsheronian molluscs and ostracods suggest that the Neogene of Central Kyzyl-Kum is represented only by Upper Pliocene sediments which are designated here the Bazil'bek formation (the former "Turan" formation of S. A. Kushnar').

This paper deals with some paleogeographic problems of Kyzyl-Kum and western Turkmenia, in the Apsheronian, and sets forth some considerations on the nature of post-Pliocene tectonic movements in central Kyzyl-Kum.

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Neogene deposits are widely developed in plainsland of western Uzbekistan where they occupy considerable areas. Their stratigraphy has not been worked out in sufficient detail, as yet; for this reason, many problems of their age and correlation are moot and in need of further study.

Neogene deposits of the Uzbekistan plains have the following characteristic features:

1) a small thickness, 600 m and possibly more only in some localities of the Bukhara trough; elsewhere, they are mostly a few tens of meters thick, seldom exceeding 100 m;

2) an incomplete stratigraphic section, with most of the Neogene missing;

3) a predominance of Pliocene deposits;

4) occurrence on an eroded Cretaceous and Paleogene surface, commonly in erosional depressions;

5) a considerable uniformity in over-all composition, accompanied by an inconsistency and variability of rocks, with arenaceous-argillaceous deposits the most common;

6) their usually undisturbed position;

7) the scarcity of organic remains, which is a reason for the controversy about their age.

It should be added that Neogene deposits are not blanket deposits, commonly being confined to more or less isolated areas; each deposit is substantially different, leading to the necessity for local stratigraphic names.

In this paper, the discussion of Neogene deposits is confined to one such area, central Kyzyl-Kum, where valuable data have been obtained in recent years as a result of field work, geologic and hydrogeologic surveying, and occasional drilling. These data lead to some important conclusions. It is believed that these conclusions will have a bearing on the study of the Neogene in adjacent provinces.

The name of central Kyzyl-Kum is applied to an area of relict highs formed by Paleozoic rocks, the Kul'dzhuk-Tau, Auminza-Tau, Aristan-Bel'-Tau, Tamdy-Tau, Bukan-Tau, etc., fringed by low undulating plains; also present here are a number of issueless troughs, the Agytmin, Karakatin, and Minbulak being the largest of them. Some authors [1, 3, 6] designate this area as southwestern Kyzyl-Kum.

Cretaceous and Paleogene deposits are widely distributed here. The youngest pre-Neogene deposits are represented by Oligocene redbeds combined into the Sarbatyr and Agytmin formations. The Sarbatyr formation (up to 64 m thick) is represented in its lower part by red to motley claystone, with the upper part consisting of gray to yellow sand with subordinate lavender clay and shell beds. Not uncommon are impressions and molds of pelecypods. Among foraminifera, N. Ye. Minakova has identified Nonion krimholzi Balachm., N. uzbekistanensis Byk., Cibicides

[1]Neogenovyye otlozheniya Tsentral'nykh Kyzyl-Kumov.

macrurus Byk., and Caucasina schischkinskyae (Sam.). Present among ostracods are Cythereis aff. jurinei Münst., Cuneocythere praesulcata Lkls., Pterygocythereis aff. impolita Mdlst., etc., The sharp difference in the lilbology and fauna, along with the local presence of gravel and pebbles at the base of the formation, suggests a break between it and the underlying Eocene claystone.

The Artymin formation is represented by a monotonous sequence of red-brown to brown claystone and siltstone, locally sandy, mostly continental, up to 150 m thick. It carries occasional peculiar fresh-water ostracods, Limnocythere kuschnari Gr., Eucypris? aggeratus Gr., E.? excisius Gr., with marine or lagunal Cytheropteron and Cytheridea in the basal horizons. The Sarabat and Agytmin formations are connected by a gradual transition. They rest conformably on underlying deposits and, in places, have been intensively dislocated along with them.

We deem it necessary to describe briefly these Oligocene deposits because, in the course of field work, they are sometimes included in the Neogene, which is of course inadmissible. There are substantial differences between Oligocene and Neogene formations of this province, in composition, occurrence, and paleontologic characteristics; careful consideration will preclude any confusion.

Neogene deposits of this province are mostly thin, being confined to isolated fairly sizable areas, where they rest almost horizontally on different Cretaceous and Paleogene units. In many places they are concealed by Anthropogene alluvium. Their exposures are present on the flanks of the Agytmin, Minbulak, and other issueless troughs in the area of Mayli-Say and Bazil'-Bek-Kuduk wells. They have been penetrated at some points by drilling done by the Uzbek Hydrogeological Trust.

Not much has been published on Neogene deposits of this area. Their most detailed description is given by S. A. Kushnar' [1, 6] who studied them during his 1936-1037 field work. Brief information based on the S. A. Kushnar' data has been communicated by B. B. Mitgarts (1948).

S. A. Kushnar' describes these deposits as thin continental sands, calcareous sands, and to a smaller extent claystone, with occasional pebbles and gravel beds. Locally present are nodules of dense marl. The presence of Driessensia polymorpha Pall. has been indicated only for upper units in the Tazbulak trough. This section is marked by its inconsistency, both laterally and vertically. S. A. Kushnar' proposed a name for it, the Turan formation, which has become fairly popular.

Inasmuch as this name, i. e., Turan formation, is used for Upper Devonian deposits of the Minusinsk trough [8], we deem it expedient to abandon it in favor of the Bazil'-Bek formation as the name for Neogene continental deposits in central Kyzyl-Kum.

We turn now to a description of some of its sections.

The southernmost exposures of the Bazil'-Bek formation occur at the south flank of the Agytmin trough where, according to A. V. Khon (1955) and to our own observations, they extend in a very narrow belt, 10 to 15 km long.

Exposed here near the Yuz-Bala well and in steep cliffs and ravine walls east of there are fairly good exposures of sediments resting horizontally either on Eocene green claystone or on the dark red Sarbatyr (basal Oligocene). The total thickness of this sequence is 34 m.

Its lower part, 13.8 m thick, is made up of drab sandstone with gravel and pebbles, gray gravel to coarse pebble conglomerate, and brown to yellow-brown claystone with gypsum crystals. Occasionally the claystone change is to mottled siltstone. Gravel and pebbles consist of Paleozoic rocks of various degrees of rounding. The following fresh-water ostracods have been observed in this section: Cyprideis littoralis (Br.), Limnocythere aff. inlauta Mdlst., L. sp., Eucypris jusbalensis Gr., E. sp., Cyprinotus agitmensis Gr., C. aff. baturini Schn., Candonella sp., Candona aff. kirgizica Mdlst., Potamocypris aff. plana Schn., Ilyocypris cf. bradyi Sars, Darwinula stevensoni (Br. et Rob.), and Characeae. The overlying interval, 17.5 m thick, is marked by a general pale-yellow color; it presents an alternation of pale-yellow and light-brown claystone carrying fine pale-yellow to gray sand grains. Also present here are thin layers of nodules of a dense light-brown to pinkish dolomite. The section is terminated with a light-brown, locally gray, cross-bedded sandstone with inclusion of Paleozoic gravel and lenses of argillaceous pebbles. The lateral inconsistency of rocks is well demonstrated in this section.

In areas adjacent to Yuz-Bala, Neogene deposits are either missing or concealed under the Anthropogene. Only the upper part of this sequence is exposed in gentle slopes of the Agytmin trough, to the west of Yuz-Bala; the same is true for the Bazil'-Bek well area, 45 km to the north, where the section is represented by brown to light-brown claystone with rare, thin intercalations of greenish-gray to gray claystone (8 m) overlain by pale-yellow to yellow clay (3 m). Numerous fresh-water ostracods have been observed here in a number of beds: Cyprideis littoralis (Br.), Limnocythere aff. inlauta Mdlst., L. dissortis Mdlst., L. aff. vara Liep., L. campestra Gr., L. kuschkanensis Gr., L. sp., Eucypris aff. concinna Schn., E.

semelosus Gr., E. samgarensis Gr., E. sp., Cyprinotus vialovi Schn., Candona rostrata Br. et Rob., Potamocypris aff. plana Schn., Ilyocypris bradyi Sars, Il. gibba (Ramd.), Il. caulatus Gr., Darwinula stevensoni (Br. et Rob.).

Drilling data show that the Bazil'-Bek formation rests in this area on an eroded surface of red Agytmin rocks. Apparently they occupy a fairly large area and extend quite far east under the cover of the Anthropogene. They were penetrated at a depth of 12 to 50 m in a borehole by the Alenda well (30 km east-southeast of Bazil'-Bek), with the Agytmin Oligocene rocks below them. Ostracods Cyprideis littoralis (Br.), Limnocythere dissortis Mdlst., L. ex gr. quadrata Mdlst., Candonella sp., and Candona angulata Müll. have been found here.

A fairly thick Bazil'-Bek section was penetrated by the borehole drilled by I. Ye. Kargin and L.I. Podlyavskaya, 25 km southwest of Alenda (Figure 1). Lying below an Anthropogene cover represented by fine-grained, yellow-gray sand on top and gravel at the bottom (the Kyzyl-Kum?), with a total thickness of 21 m, is the following section (from top to bottom):

4. Gray sandstone with ochre-colored spots, fine-grained, micaceous; cement calcareous-argillaceous; thickness 15.5m.

5. Yellow to gray claystone with ochre-colored spots of iron hydroxide and black bloom of manganese oxides; intercalations of gray sandstone; thickness 4.6 m.

6-7. Yellow claystone, locally gray with black manganese oxide spots; a thin (0.3 m) yellow-gray sandstone layer occurs in about the middle of it. Somewhat above this sandstone, at a depth of 50 m, isolated salt water ostracods Cythere ex gr. variabiletuberculata Schw. have been observed along with rare remains of fresh-water ostracods; thickness, 19.6 m.

8. Gray sandstone with ochre-colored spots, fine-grained, micaceous; argillaceous cement; thickness, 7.1 m.

9. Pale-yellow clay, unstratified, with black manganese bloom; 2.2 m thick.

10-11. Sandstone, brown-gray, fine-grained on top, with inclusions of small pebbles; changing downward to brown fine-grained sand with claystone pebbles and gravel; 19.5 m thick.

12. The agytmin formation. Clayston, red-brown to red with black manganese bloom, unstratified, with intercalations and lenses of gray sandstone at a depth of 135 m. Typical Agytmin fresh-water ostracods occur; e.g., Limnocythere kuschnari, occur at a depth of 55.5 m.

Beds 4-11 constitute the Bazil'-Bek formation, up to 68.5 m thick. It rests on the eroded surface of the Agytmin formation.

The Bazil'-Bek formation occupies a fairly large area adjacent to the Mynbulak trough, where it is exposed in isolated spots and is often covered by alluvium. It extends from the southeast end of the trough toward the Mayli-Say well. According to K.K. Pyatkov, it rests either on green Eocene clays or on Oligocene red clay. Its thickness (probably incomplete) is about 10 m. It is represented by silty clay (greenish-gray below; gray-yellow to yellow above) alternating with yellow siltstone, friable sandstone, and siliceous quartz gravel beds. Some layers carry fresh-water ostracods, Cyprideis littoralis (Br.), C. torosa (J.), Limnocythere ex gr. dissortis Mdlst., and Candonella ex gr. schubinae Mdlst.

According to S.A. Kushnar' [6], these deposits are present in the flanks of minor issueless troughs located south of the Mayli-Say well: the Tuz-Bulak, Kara-Kuduk, and Tamdy-Kuduk, where they are represented by very variable arenaceous-argillaceous sediments, carrying pebble beds. Their thickness in the Kara-Kuduk trough is 16.8 m. Present in pebble beds of the upper part of these deposits in the Tuz-Bulak trough are numerous Dreissensia polymorpha Pall.

Data on Neogene deposits in the area northwest of the Bukan-Tau Mountains are extremely scarce because of the widespread cover of sand. A.V. Pakhomova (the Uzbek Hydrogeological Expedition, 1954) discovered poorly consolidated alluvial sand and friable brown-gray sandstone with thin intercalations of light-brown to brownish claystone in a number of diggings and hand-bored holes. At some points; the incomplete thickness of these deposits is 15 to 20 m. They rest on an eroded surface of green Paleogene claystone. A number of samples contained locally numerous fresh-water ostracods, Cyprideis littoralis (Br.), C. torosa (J.), Limnocythere aff. vara Liep., L. septuosa Gr., L. mubarekensis Cr., L. dengiskulensis Gr., L. sp., ilyocypris, Candonella. The over-all aspect of these sediments, along with their microfauna, suggests that they belong to the same Bazil'Bek formation.

Deposits of this formation have also been found by M.T. Burak on the south flank of the Karakatin trough, approximately 20 km southwest of Chingel'da village. Here, they are represented by a thin sequence of green-gray silty clays resting on green Eocene clay and overlain by Anthropogene gravel. Present in these clays are fresh-water ostracods Cyprideis littoralis (Br.), C. torosa (J.), Limnocythere inderica Schar., L. campestra Gr., L. aff. impressa Gr., Ilyocypris, Darwinula.

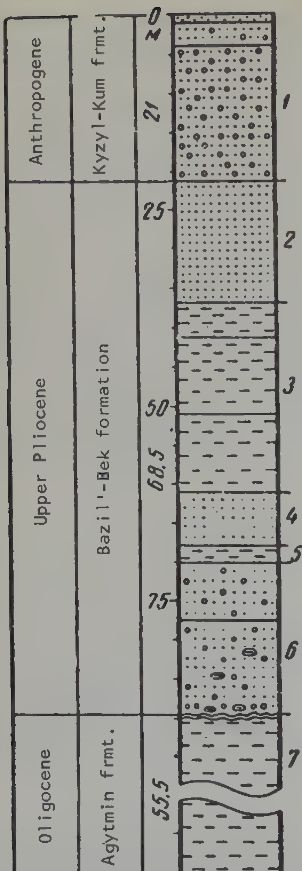

FIGURE 1. Bazil'-Bek section penetrated by borehole 25 km southwest of Alenda.

1 - gravel and sand; 2 - sandstone, gray, fine-grained; 3 - yellow to yellow-gray claystone with rare intercalations of gray sand; Cythere ex gr. variabiletuberculata occurring at a depth of 50 m; 4 - gray, fine-grained sandstone; 5 - yellow claystone; 6 - sandstone brown-gray changing to brown, medium- to mixed-grained with gravel and pebbles of claystone; 7 - red-brown claystone with Limnocythere kuschnari.

Of importance is a section at the northwestern end of the Minbulak trough (5 km east of Minbulak village). Here, the Bazil'-Bek formation, 33.3 m thick, rests on an erosional cut in green Eocene clay. Its lower part consists of green-gray clay with intercalations of sandstone carrying moderately well-rounded pebbles of quartz, chert, and metasedimentary Paleozoic rocks as well as fragments of Lower Paleogene claystone and marl; its middle interval carries a 1.5 m thick bed of white fine-grained marl. This member is 6.15 m thick. It is followed by buff to light-brown clay, silty in the upper part and carrying thin silty sand beds; approximately in the middle of it there is a thin layer of flat pebbles of gray and dense upper Eocene sandstone. The lower member contains fresh-water Apsheronian molluscs Corbicula iluminalis Müll. var apscheronica Andrus. and Melanoides apscheronica Andrus.

A number of beds throughout the section carry ostacods Cythereis pseudoconvexa Liv., Cytherissa ex gr. bogatschovi Liv., Limnocythere tenuireticulata Suz., L. ex gr. pomosa Mdlst., L. inopinata (Baird), L. ex gr. vara Liep., L. vitrea Gr., Cyprideis littoralis (Br.), C. torosa (J.), Eucypris paululus Gr., Zonocypris limestis Gr., Canlonella schubinae Mdlst., Ilyocypris bradyi Sars, Il. gibba (Ramd.), Darwinula sp. Molluscs and ostracods Cytheris pseudoconvexa, Cytherissa ex gr. bogatschovi (brackish water forms), and Limnocythere tenuireticulata suggest that these deposits are Apsheronian. The mixed character of the fauna, with brackish water forms present along with the predominant fresh-water ones, together with their ecology and the nature of the sediments, suggest that here we deal with deltaic deposits belonging to a channel of a large river debouching into an Apsheronian basin [4].

Apsheronian deposits have been discovered recently also south of the Minbulak trough, in the areas of Chokolak (20 km south of Minbulak village) and It-Ashchi wells (20 km southwest of Chokolak). Here, according to P.M. Sveshnikov [7] and L.A. Ostrovskiy [4], they are represented by a thin member of gray to green-gray claystone, commonly calcareous, with intercalations of friable sand and sandstone, resting on an eroded surface of green Eocene clay, and possibly even on Cretaceous rocks; in most places, they are covered by recent deposits.

The Apsheronian age of these rocks has been determined by the presence of marine (more precisely, brackish) ostracods Cythereis pseudoconvexa Liv., Cytherissa cascusa Mdlst., Cythere cellula Liv., and C. ex gr. variabiletuberculata Schw. The presence of a large number of fresh-water forms (Cyprideis littoralis, C. torosa, Ilyocypris bradyi, Candona angulata, and Limnocythere ex gr. vara) suggests a considerable freshening of the basin. Judging from its position, this basin was connected in some way with the Khorezmiyskoye Lake which occupied the lower Amu-Dar'ya depression.

It appears from these descriptions that sediments of the Bazil'-Bek formation within central Kyzyl-Kum were of the same type, on the whole, despite their differences. The similar conditions of occurrence, along with a similar fauna of fresh-water ostracods, lends substance to the belief that we deal here with contemporaneous deposits.

The age of these deposits was determined by S.A. Kushnar' as Neogene to Early Quaternary ($Ng + Q_1$) and the assumption was that they

include all of the Neogene, down to its base. B. B. Mitgarts (1948) found it possible to raise the lower age boundary and tentatively assume the age of these deposits to be Pliocene to early Quaternary. It should be stressed that neither S. A. Kushnar' nor B. B. Mitgarts had any paleontologic data at their disposal and were fully justified in their opinions on the basis of what they knew.

Paleontologic data gathered in the last few years pinpoint the age of the Bazil'-Bek formation. These data are as follows: 1) typical Apsheronian molluscs (Corbicula fluminalis var. apscheronica, Melanoides apscheronica) and brackish and fresh-water ostracods (Cythereis pseudoconvexa, Cythere ex gr. variabiletuberculata, Limnocythere tenuireticulata, etc.) found in the area of the Minbulak trough; 2) brackish-water ostracods, Cythere ex gr. variabiletuberculata from a borehole south of the Aristan-Bel'-Tau highs (southwest of Alenda well).

These fossil remains fix the age of the Bazil'-Bek formation as Apsheronian (Upper Pliocene).[2]

Throughout most of the central Kyzyl-Kum area, the conditions of formation of the Bazil'-Bek beds were undoubtedly continental. S. A. Kushnar' believes [6] that lacustrine basins were of prime importance in the accumulation of these deposits. It is likely that they were located in a vast alluvial plain whose relief was emphasized by isolated low ridges of Paleozoic rocks. It was from these ridges that intermittent streams carried the gravel so prominent in the Bazil'-Bek formation of this area. The hydrographic network apparently was complexly dendritic but the general direction of the stream flow was almost latitudinal.

It appears that at least two river courses, rather than a single one, were operative here, at that time. The ancient valley of the Syr-Dar'ya apparently was located in the north (Mayli-Say well and Bukan-Tau Mountains areas); its delta (at least a branch of it) is suggested in the northwest of the Minbulak trough. In the south (Yuz-Bala well area) was another stream probably belonging to the system of the ancient Zeravshan.

Two groups of lacustrine basins can be projected on the basis of the areal distribution of the Bazil'-Bek sediments.One group was located south of the Aristan-Bel'-Tau and Auminza-Tau Mountains, in the area of wells Alenda, Bazil'-Bek, and Oydyn-Bulak. The other group was located west of Tamdy, in the area of the Mayli-Say well and the present day issueless troughs of Tamdy-Kuduk, Tuz-Bulak, and Kara-Kuduk.

The existence of brackish-water organisms (Cythere) south of the Aristan-Bel'-Tau suggests an inconsistent salinity of lacustrine basins, as well as some connection, be it ever so remote and intricate, with the Apsheronian Caspian basin.

The data on hand suggest a configuration of the Apsheronian basin in this part of Uzbekistan and western Turkmenia, as indicated in Figure 2.

1. The main Apsheronian basin, the most saline, was located in the Caspian area to the west. An embayment of that basin possibly penetrated deep into the low Kara-Kum, far to the east.[3]

2. East of there was a group of brackish, major lakes, including the Sarykamysh Khorezmian (lower course of the Amu-Dar'ya), Aral, and west Kyzyl-Kum. There salinity decreased eastward, permitting the existence of a poor brackish-water macro- and microfauna.

3. Still farther east, in the central Kyzyl-Kum area, the vast alluvial plain was strewn with numerous small, mostly shallow fresh-water lakes interconnected by channels. From time to time, because of evaporation, the salinity of these basins would rise to create conditions relatively favorable for a brackish-water microfauna brought in by birds and even by wind. The possibility cannot be ruled out, however, that direct connection existed from time to time between these and the group of larger basins to the west.

The absence here of older Neogene and Eocene sediments, along with the position of the Bazil'-Bek formation on an eroded surface(locally in erosional depressions) of various older deposits, with the Agytmin the youngest among them, all points to a very poor development of sedimentary processes in the Miocene and Lower and Middle Pliocene. At that time, erosion rather than sedimentation was the dominant factor. The nature of geologic process changed only in the Late Pliocene, at which time the Bazil'-Bek formation began to be deposited.

The distribution of these sediments and their sediments and their features suggest that in the Late Pliocene, unlike now, this province was irrigated to a considerable extent by an extensive system of rivers and lakes. The thinness of sediments, seldom exceeding a few tens of

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[2]It is not impossible that the base of the Bazil'-Bek formation, at least locally, is Akchagylian. Russian Editorial Board.

[3]In Figure 2, the east boundary of the Apsheronian (Caspian) basin has been drawn from data published in "Geologiya S. S. S. R., " t. 22 [2]. Gostoptekhizdat, 1957.

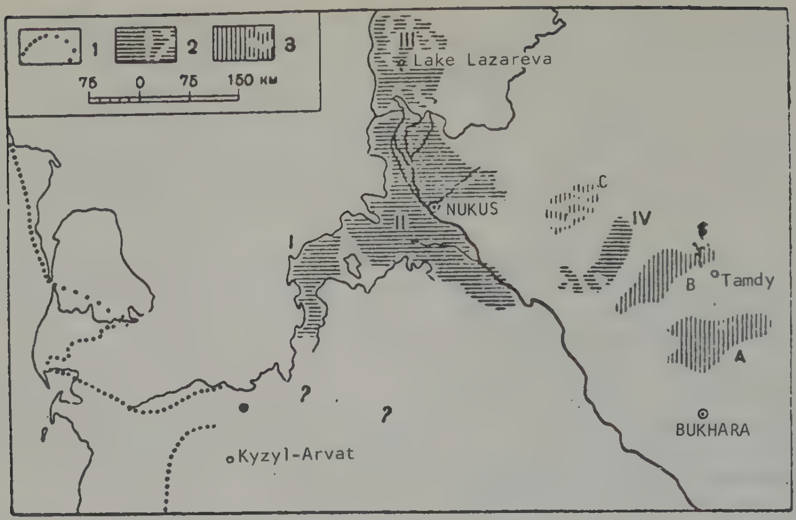

FIGURE 2. Approximate distribution of Apsheronian basins in the western part of the Uzbek S.S.R. and adjacent provinces of western Turkmenia.

1 - major lacustrine basins: I) Sarykamysh; II) Khorezmian; III) Aral; IV) West Kyzyl'-Kum; 2 - areas of saltwater basins; 3 - areas of chiefly fresh-water lakes: A) south of Aristan-Bel'-Tau; B) west of Tamdy; C) northwest of Bukan-Tau.

meters, (usually not over 15 or 20 m), reflects tectonic conditions under which subsidence was very slight, in any event considerably less intensive than in adjacent provinces.

More specifically, central Kyzyl-Kum border the southeastern or Dzhizak Kyzyl-Kum which contains an Upper Pliocene trough where these sediments are fairly thick. A test drilled by the Uzbek Hydrogeological Trust at the Yergash-Kuduk well (north of the Aristan-Bel'-Tau Mountains) penetrated almost 250 m of yellow-gray to gray fine-grained friable sandstone with definitely subordinate thin claystone and fairly common gypsum pockets; the base of this section lies at 276 m. The similar Bukhara trough is located to the south. A correlation of these data suggests that the Late Pliocene central Kyzyl-Kum presented a relatively uplifted segment within the Uzbekistan plain.

As already pointed out, the Bazil'-Bek rocks in central Kyzyl-Kum are almost horizontal, resting unconformably on various older pre-Neogene deposits. On the whole, these latter beds are quite flat, almost horizontal; locally, however, they are strongly dislocated and are steeply dipping. Such disturbances have been observed not only in the immediate vicinity of Paleozoic outcrops, which in addition may be thrust over the Cretaceous and Paleogene, but a fairly long distance away from them, as well (Agytmin trough). Considering that the Bazil'-Bek formation has not been involved in the folding, it may be assumed that the folding had been the most intensive in the pre-Late Pliocene. Whether it occurred at the close of the Oligocene or later (pre-Akchagyl movements), is hard to say.

It is hardly possible to find a satisfactory answer to this question within central Kyzyl-Kum, at the present time, because of the absence of Neogene beds older than the Upper Pliocene.

It is not impossible that these disturbances were associated with pre-Akchagyl movements which resulted in considerable subsidences in the Volga-Caspian province and in penetration by the Akchagyl basin as far as the present delta of the Amu-Dar'ya [5].

As a result of pre-Pliocene disturbances, highs of Paleozoic rock became prominent in the central Kyzyl-Kum relief. This is confirmed by the presence of coarse clastic material in the Bazil'-Bek formation.

Tectonic movements continued after the deposition of that formation, i.e., in the Anthropogene. They were expressed on the whole in a further uplift of Paleozoic ridges. At the present time, there are no data to the effect that these movements have tilted the Bazil'-Bek beds. There is a single mention by S. A. Kushnar' that Neogene deposits in the Tamdy-Kuduk trough dip gently to the south. However, locally

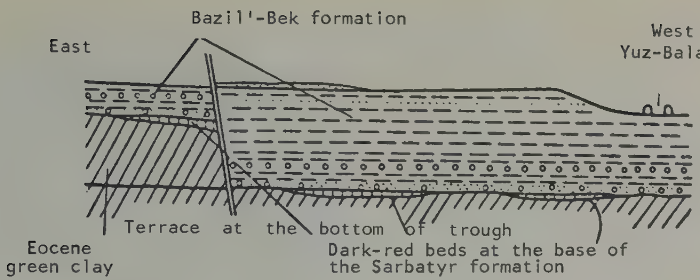

FIGURE 3. Diagrammatic sketch of a fault in the Bazil'-Bek formation east of Yuz-Bala (south flank of the Agytmin trough)

fairly large faults, originated in some places. A small fault with a throw of several meters has been observed in the Bazil'-Bek formation, several hundred meters east of Yuz-Bala (Figure 3). Apparently connected with the same post-Pliocene tectonic movements is a large fault running along the northeastern flank of the Agytmin trough. According to A. V. Khon (1955), it is no less than 40 km long, with the stratigraphic throw locally as much as 100 m. Beyond the trough proper and north of it (8 km northwest of the Kosh-Kuduk well), almost horizontal Upper Pliocene beds have been observed along the fault line, displaced vertically to contact steeply dipping (∠75-90°) Turonian sandstone. This fact suggest the magnitude of post-Pliocene movements in the central Kyzyl-Kum province.

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ANALCITE- AND ZEOLITE-BEARING SEDIMENTARY ROCKS OF TUVA[1]

by

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Zeolite-bearing and essentially analcitic sedimentary rocks, formerly regarded as a rarity, have been found by this author to have a wide regional distribution in association with rocks of various ages from Paleozoic to Recent. Sedimentary rocks carrying authigenic laumontite (zeolite) and analcite, as well as essentially analcitic chemogenic rocks, were studied by the author from Middle Devonian and Lower Carboniferous sections in the Tuva Autonomous Oblast'.

In addition to the personally recorded data, this paper summarizes published data on the zeolite distribution in deposits of various ages.

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1. A Brief Petrographic-Mineral Description of Analcite- and Zeolite-Bearing Sedimentary Rocks of Tuva

We have discovered authigenic analcite and zeolite-laumontite in the Tuva Autonomous Oblast' in Middle Devonian and Lower Carboniferous sedimentary formations, where laumontite locally is one of the principal components of D_2 sandstone cement and analcite is widely developed in the cement of Lower Carboniferous terrigenous deposits; these rocks also contain intercalations and lenses of chemogenic, essentially analcitic sedimentary rocks.

a) Middle Devonian Laumontite-bearing Sandstones of Tuva

Middle Devonian deposits of Tuva are fairly widely developed, transgressive on the $S_2 - D_1$ sedimentary-volcanic sequence [15]. Fossiliferous Eiffel and Givetian stages have been identified among them. The Eiffel stage is made up of volcanic-terrigenous formations; the Givetian consists of the terrigenous beds which have been divided [9] into three formations from top to bottom: the Atakshil', Ilemorovo, and Uyuk.

An intensive authigenic development of zeolite is associated with the Atakshil' formation (Atakshil', Atyktash Rivers, etc.) consisting of red (less commonly gray) sandstone and occasional gravel beds and conglomerate. Seen under the microscope, these sandstones are uneven-grained, less commonly well sorted; intergranular cement is massive, and in places developed at the grain contacts; clastic material, polymictic. Grains are angular to fairly well-founded, represented chiefly by quartz (35 to 70%), feldspar (10 to 34%), chalcedony, fragments of extrusive rocks with felsitic, micropoikilitic, and other groundmass (7 to 38%) textures; and consist of silty pelite, pelite, and quartzite. Always present are some of the epidote group minerals, garnet, zircon, apatite, tourmaline, rutile, sphene, spinel, ilmenite, magnetite, chromite, and occasionally hornblende, pyroxene, and corundum; the sandstone cement is calcite and laumontite. Figure 1 illustrates the distribution of laumontite in sandstone; brick-red laumontite fills up the interstices between clastic grains.

Other authigenic formations, besides calcite and laumontite, are biotite, muscovite, chlorite, barite, and octahedrite. Extrusive clastic material has been partially decomposed with the formation of chlorite and sericite. No zeolites formed from redeposited extrusives have been observed. This is true even for direct contacts between laumontite of the pore space in sandstones and extrusive fragments.

b) Essentially Analcitic and Analcite-bearing Lower Carboniferous Sedimentary Rocks of Tuva

Authigenic analcite is widely distributed in Lower Carboniferous sedimentary rocks of Tuva. Present among them are chemogenic, essentially analcitic and terrigenous analcite-bearing rocks represented by fine-grained motley sandstone, tuffaceous sandstone, tuff, shale, and phosphate-calcareous siltstone. They rest transgressively, with an angular unconformity, on Devonian and older deposits [15].

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[1]Anal'tsim- i tseolitsoderzhashchiye osadochnyye porody Tuvy.

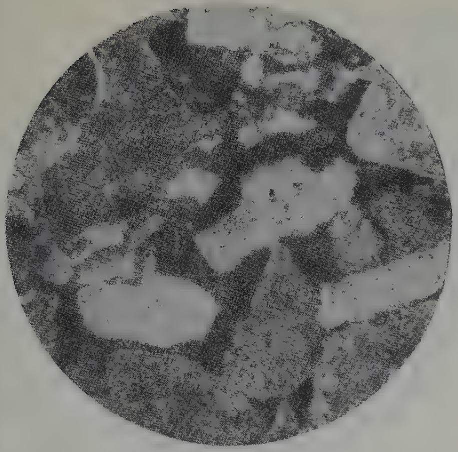

FIGURE 1. Brown laumontite in the sandstone cement

Black, laumontite. Without analyser; 60 X.

Analcite cement in them has been observed at many points on either side of the Yenisey River.

Analcitic proper and tuffaceous sandstone are polymictic, interbedded with tuffs cemented with analcite. Members carrying analcitic rocks are as much as 100 m thick or more. Usually, these are fine-grained rocks, thin-bedded, in places cross-stratified, pink-gray to brown, displaying thin banding caused by laminae of ash particles replaced by analcite, alternating with sand and silt partings (Figure 2).

Analcitic sandstone and tuffaceous sandstone are polymictic, with intergranular to massive cement. Clastic material is represented by quartz, feldspars, fragments of volcanic glass, extrusives, and quartzite, with a small amount of zircon, garnet, sphene, and ore minerals. The massive cement consists of analcite and calcite, with analcite commonly predominant. Analcite in cement forms masses without any structural features; immersed in it are grains of quartz, feldspar, and other clastic material. Present in analcite are occasional ash particles recognizable by their form but fully replaced by analcite.

As in tuff, ash particles replaced by analcite are fringed by a fine-scaled mica aggregate. It has been observed that, in total replacement of ash particles by analcite, redeposited extrusive fragments consisting of brown glass with plagioclase microliths are not zeolitized but remain unaltered.

The cement of analcitic sandstone and tuffaceous sandstone often carries authigenic fluorite which locally becomes a principal cementing mineral, along with analcite. Present in sandstone are thin layers virtually fluorite-free, rhythmically alternating with those rich in fluorite.

Described below are some interesting morphologic relationships between analcite, cryptocrystalline zeolites, and fluorites, all cementing the clastic material of sandstones. Present at the base of an analcitic sandstone bed are concentrically zoned fluorite-analcite nodules, up to 3 or 4 cm in size. Their central part consists of massive analcite, and the periphery consists of a rhythmic alternation of analcitic and fluoritic zones.

Larger nodules are joined with one another; then the fluorite enriched zones, also rhythmic, bend about these ellipsoidal nodules, straighten out, and gradually change to more or less rectilinear layers. The thickness of layers strongly enriched in fluorite is a few millimeters.

The amount of clastic material in the lower part of such a bed is quite small, while authigenic minerals, fluorite and analcite (with small additions of calcite and collophane), are very prominent; because of that, the rock in these segments can be assumed to be chemogenic.

Higher in the section, the amount of clastic material increases gradually, and the rock as a whole is identified as sandstone with a massive fluorite-analcitic cement. It appears that nodules of aluminosilicate colloids, rhythmically enriched in fluorite, were formed here prior to the deposition of clastic material. Figure 3 shows a rhythmic deposition of fluorite on the analcite nodule. Also present are micronodules of collophane and calcite with an annular fringe of analcite; other micro-nodules are pure analcite.

On the left bank of Biy-Khem River (basin of Taps River), analcitic sandstone is associated with sandstone and conglomerate showing a well-expressed regeneration of clastic grains of quartz and feldspar. In the upper half of the section, having a total thickness of about 100 m, analcite is a consistent component of the cement in sandstone and conglomerate; in the lower part, clastic grains of quartz and feldspar have been regenerated. Analcitic sandstone and those with a superimposed regeneration are quite similar macroscopically; they both are fine- to coarse-grained, of different hues of red, with polymictic clastic material. Under the microscope, they show a fine-grained texture, with porous to massive cement. Their clastic material consists of quartz, feldspars, quartzite, extrusives with various groundmass textures, and associated epidote, garnet, zircon, apatite, ore minerals, and tourmaline; ash material has not been observed.

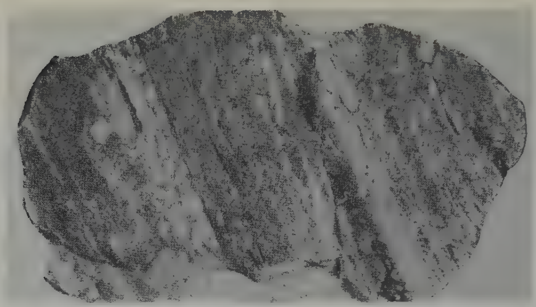

FIGURE 2. Tuffaceous sandstone.

Light-colored band lenses, analcitic tuffaceous sandstone; 4/5 natural size.

Clastic material is covered with a film ("shirt") or iron hydroxides.

The cement in analcitic sandstone is analcite and coarse-grained calcite (recrystallized), with occasional crystals of dolomite and incipient chlorite scales developed on the edge of clastic grains. Iron hydroxides are diffused in the cement; in addition, they form a thin film on clastic grains. Conglomerates are red and consist of large to medium-size pebbles. Their texture is psephitic; the cement is psammitic. The pebbles consist of andesite porphyrites, quartz porphyry with a felsitic groundmass texture, sandstone, micropegmatite, feldspar, and quartz.

The pebbles are cemented by polymictic psammitic material which, in turn, is cemented with authigenic analcite and calcite, filling the pore space. Present instead of analcite in the conglomerate and sandstone cement of the lower part of this section are regenerated clastic grains of quartz and feldspars which show their original outlines because of the "ferruginous shirt" preserved underneath the newly-formed fringes. The regenerated portion of quartz and feldspar grains is quite transparent and free of any inclusions.

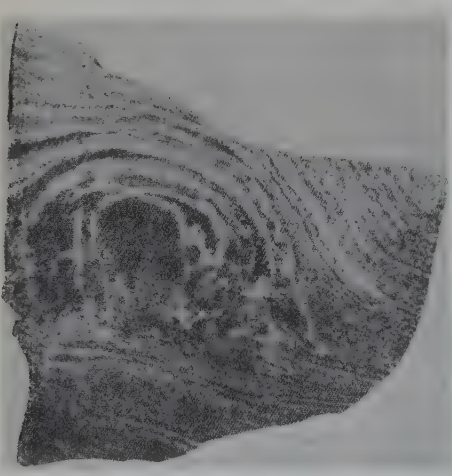

FIGURE 3. Rhythmic deposition of fluorite on an analcite nodule.

Black, fluorite, natural size.

Analcitic tuffs has been observed among Lower Carboniferous deposits (Mt. Kherbes, Suglug-Khem River). These are rocks with analcite replacing the ash particles; they change gradually to analcitic and tuffaceous sandstones, and alternate with them.

Macroscopically, dense, banded, pink-gray analcitic tuffs alternate with green-gray sandstone carrying a very small amount of ash material. These tuff beds are 3 to 3.5 m thick.

Seen under the microscope, these tuffs exhibit a crystalline vitroclastic, psammitic texture, with an ash-type cement. Typical of the ash particles is their sickle-like to forked shape; they have been fully replaced by analcite which forms pseudomorphs on these particles, fringed by micaceous aggregates. Present in small amounts in tuffaceous layers is psammitic material (quartz, feldspars, quartzites, and locally calcite in spots). Shown in Figure 4, is tuff with its ash particles replaced by analcite.

Essentially analcitic sedimentary rocks have been observed among analcitic tuffaceous sandstone and locally in sequences of analcitic sandstone and conglomerate.

The previously described [4] analcitic sedimentary rocks (Mt. Ak-Tag) occur in conformable beds (5 to 6 m thick) among phosphate, analcitic, and calcareous sandstone and tuffaceous sandstone (along with transitional beds). Visible in them to the naked eye are analcite crystals oriented parallel to the stratification, and a tough cementing substance with a plano-conchoidal break. Present locally along the

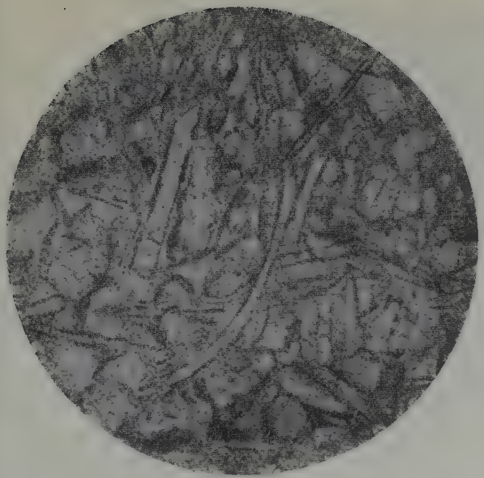

FIGURE 4. Tuff. Ash particles replaced by analcite.

160 X; without analyzer.

bedding planes are carbonized plant remains with a higher concentration of coarser analcite crystals around them.

The rock texture is porphyroblastic, because of the analcite crystals (Figures 5 and 6); the cement is fine-grained to cryptocrystalline, with an unevenly-banded texture. The amount of clastic material is small, increasing locally, but with the same cryptocrystalline cement everywhere.

This cryptocrystalline cement consists of zeolites: laumonite and epistilbite (?). Essentially analcitic rocks on the left bank of Biy-Khem River (the Taps basin) occur in a lens between two conglomerate beds with analcitic cement; it is up to 2.5 m thick and is traceable only for about 15 m.

Macroscopically, this rock is psammitic, medium- to coarse-grained, 90% closely packed analcite crystals, equal to 1 mm, and 10% cement material; in isolated segments, analcite crystals account for almost 100% of the total rock. The structure is psedo-oölitic, formed by spherical to ellipsoidal analcite grains (Figure 7), with general regular grain cross-sections. Some analcite crystals contain clastic grains of quartz, and feldspars, and rare apatite grains; it looks as though clastic material was immersed in aluminosilicate gel, subsequently crystallized as analcite. The interstices between these crystals are occupied by iron hydroxides. No pyroclastic material has been observed. In segments of monomineral analcite rock, analcite crystals appear to be arranged in "tightest packing", in the center of crystals are concentrations of finely dispersed orange iron hydroxides; their contact faces show carbonate and aggregates of thin-scaled mica. In isolated segments, bitumens produce drops of a dark brown petroleum-like substance.

These essentially analcitic rocks were chemically analysed (Table 1). A conversion of the chemical analysis [1] to minerals shows 34.34% analcite and 62.69% cryptocrystalline cement, which compares well with observation results.

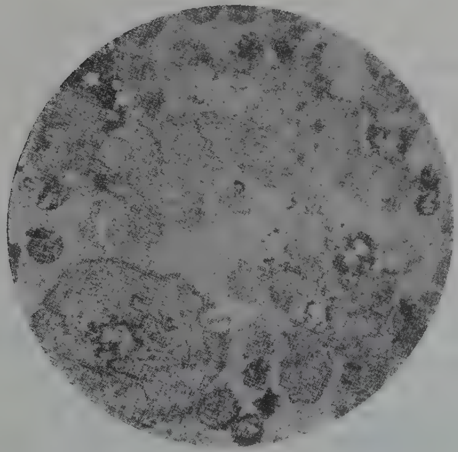

FIGURE 5. Essentially analcitic rock with porphyroblasts of analcite (right bank of Yenisey; Mt. Ak-Tag).

25 X; without analyzer.

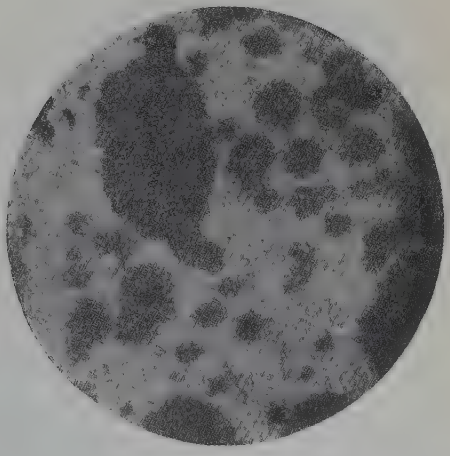

FIGURE 6. Essentially analcitic rock with porphyroblasts of analcite (same area).

25 X; black, analcite; crossed Nicols.

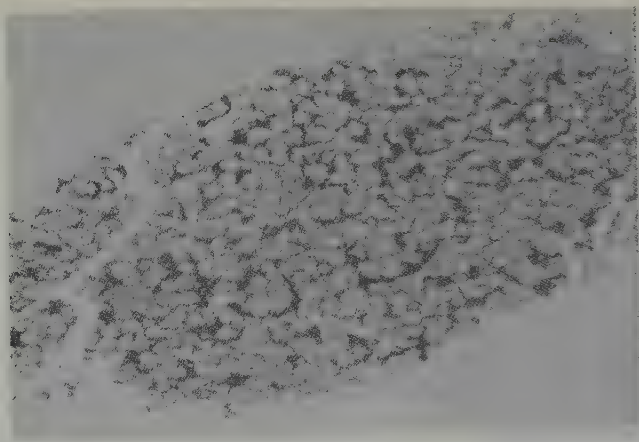

FIGURE 7. Essentially analcitic rock (Taps River basin).

Thin section; 5 X.

As shown by the conversion of chemical analysis [2], the analcite content does not exceed 52%, compared to 90% and more as visually determined. This is because a considerable amount of clastic material is cemented within the analcite crystals.

Data cited in Table 1 are very close to those obtained by W. Bradley [25] for analcite rocks from the Green River formation. According to Bradley, the computed mineral composition for the Green River formation is as follows: analcite, 65.84%; chalcedony, quartz, and opal, 15.66%; sanidine and orthoclase, 3.71%; limonite, 11.3%; pyrite, 2.73%; calcite, 0.69%. The texture of rocks described by W. Bradley, too, is similar to ours.

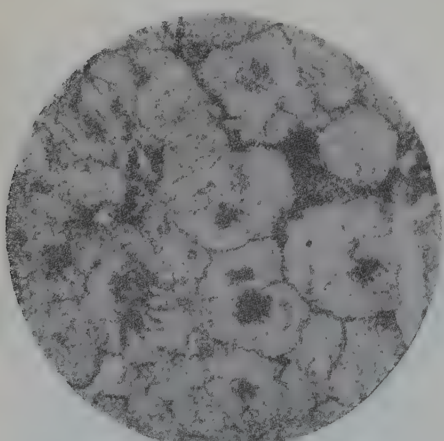

FIGURE 8. Essentially analcite rock (Taps River basin).

Black, nodules of analcite in the center of iron hydroxide crystals. 10 X; without analyzer.

II. Analcite, Zeolites, and their Principal Authigenic Associates in Sedimentary Rocks of Tuva

Analcite in sedimentary rocks has been observed in the following forms: a) crystals; b) pseudomorphs on ash particles; c) xenomorphic grains filling up the space between clastic grains.

Analcite crystals from analcitic rocks and analcitic conglomerates have their standard tetragonal-trioctahedral form and are occasionally 3 mm in size; under the microscope, they show hexagonal to octagonal sections. They are grayish-white to yellowish, with a vitreous luster. Small crystals are transparent. Hardness, 5 to 5.5. A rather poor cleavage is present. Analcite is isotropic, with $N = 1.485 \pm 0.002$. Its crystals commonly contain inclusions of clastic grains of quartz, feldspars, and some unidentifiable minerals.

Analcite has a positive reaction for SiO_2; in a closed tube, it gives off water and is readily fused to a white, bubbly glass.

An X-ray analysis by Ye. P. Sokolova (VSEGEI) shows that this mineral is indeed analcite.

Laumontite is white to pink and brick-red; under the microscope, it is colorless to red-brown. Studied in reflected light, with high magnifications, intensively colored laumontite

Table 1

Chemical analysis of essentially analcitic rocks[1]

Components	Right bank of the Yenisey [1]		Conversion to minerals					
	Weight per cent	Molecular amount	Fluorite	Analcite	Ilmenite	Magnetite	Hematite	Cement
SiO_2	66,56	1108	—	312	—	—	—	890
TiO_2	0,11	001	—	—	001	—	—	
Al_2O_3	15,77	155	—	078	—	—	—	077
Fe_2O_3	0,91	006	—	—	—	003	003	—
FeO	0,28	004	—	—	001	003	—	—
MnO	0,04	0006	—	—	—	—	—	0006
MgO	0,28	006	—	—	—	—	—	006
CaO	1,36	024	010	—	—	—	—	014
Na_2O	4,87	078	—	078	—	—	—	—
K_2O	4,26	046	—	—	—	—	—	046
H_2O-105°; Losses in heating	0,27; 4,45	287	—	156	—	—	—	131
CO_2								
F	0.60	021	021	—	—	—	—	—
P_2O_5								
FeS_2								
Bituminous material								
Total	100,21	—	1,64%	34,34%	0,15%	0,70%	0,48	62,69%
—O=F_2	—0,22							
	99,99							

Components	Left bank of the Yenisey [2]		Conversion to minerals								Bradley's data [25]
	Weight per cent	Molecular amounts	Analcite	Apatite	Fluorite	Hematite	Ilmenite	Dolomite	Magnetite	Clastics, etc.	Weight per cent
SiO_2	50,04	833	468	—	—	—	—	—	—	365	55,54
TiO_2	0,30	004	—	—	—	—	004	—		—	—
Al_2O_3	17,25	170	117							053	15,78
Fe_2O_3	10,36	065				037			028	—	4,54
FeO	2,33	032					004		028	—	
MnO	0,05	0007								0007	—
MgO	1,59	040						016		024	0,63
CaO	1,44	026		001				015		007	0,36
Na_2O	6,94	112	117		003					—	8,63
K_2O	0,54	005								—	1,16
H_2O-105°; Losses in heating	0,97; 6,63	423	234							—	0,05
CO_2	1,33	031						031		129	—
F	0,16	005									8,96
P_2O_5	0,06	0004		0004	005						0,29
FeS_2	—	—									—
											2,74
Bituminous material	—	—									3,94
Total	100,01	—	51,50%	0,13%	0,39%	5,91%	0,61%	5,72%	6,48	29,26%	99,61
—O=F_2	—0,07										
	99,94										

[1] V. M. Kovyazina, Analyst.

shows inclusions of fine hematite scales, with red internal reflections which determine the color. In addition, it contains numerous fine liquid inclusions; these bubbles are especially conspicuous in immersion preparations.

This mineral has a biaxial negative form, with an average 2V; interference color, white; extinction oblique; Sp $\gamma = 35°$, $\gamma' = 1.526 \pm 0.002$; $\alpha' = 1.513 \pm 0.002$; $\gamma' - \alpha' = 0.013$. Cleavage well expressed along (010); less perfect prismatic cleavage.

Laumontite forms a SiO_2 gel with HCl; in a closed tube, it gives off water; fuses readily to white, bubbly, very brittle glass; isotropic after fusing: $N = 1.532 \pm 0.002$.

The principal elements in laumontite as determined spectroscopically are silicon, aluminum, and calcium, with some Mg (about 0.03%), Fe (0.03-0.1%), Mn (0.001-0.003%), and Ba (about 0.03%) Ye. Ya. Smirnova, Anayst). Ye. P. Sokolova (VSEGEI) performed an X-ray analysis on brick-red laumontite [5], which proved its identity with standard laumontite.

Epistilbite (?) has been identified only in analcitic rocks from the right bank of the Yenisey (Mt. Ak-Tag). It forms fringes about idiomorphic laumontite grains and constitutes a principal component of the cryptocrystalline cement of these rocks. The following data were obtained for epistilbite: biaxial, positive, extinction angle $c\gamma = 9°$; average 2V; $\gamma = 1.517 \pm 0.002$; $\alpha = 1.506 \pm 0.002$; $\gamma - \alpha = 1.011$; this zeolite has not been positively identified because of the impossibility of isolating a monomineral fraction for additional study.

Principal authigenic associates of analcite and zeolites in the Tuva sedimentary rocks are fluorite, collophane, calcite, dolomite, and, locally, sulfides: sphalerite, galena, pyrite, and chalcopyrite.

Fluorite is widely distributed in Lower Carboniferous deposits where it locally forms rich accumulations in cement, in association with analcite, cryptocrystalline zeolites, collophane, and calcite. The rhythmic deposition of fluorite in this association has been described above.

Fluorite has been deposited in the cement in Xenomorphic grains filling the interstices in clastic material, and in purple cubic crystals along the bedding planes. In places, its coloring is very weak, and the color distribution is uneven, i. e., more intensive near the edges and at the tops of crystals. Occasionally, the coloring is zoned; $N = 1.434$.

Identified in purple fluorite by spectrographic analysis are Ca (10%), Sr (0.01-0.1%), and Ba (0.01%) (VSEGEI; L. I. Denisenko, Analyst).

Collophane is a cryptocrystalline calcium phosphate of the apatite group, developed in the cement of Lower Carboniferous sedimentary rocks. Pure collophane grains, free of inclusions, are yellowish-gray, translucent; luster, waxy to greasy.

Seen under the microscope, collophane is yellowish to slightly yellow-green; crossed Nicols reveal isotropic sectors as well as slightly birefringent ones, with small polarization areas. Refractive index for isotropic collophane, $N = 1.615 \pm 0.002$.

X-ray analysis (VSEGEI; Ye. P. Sokolova, Analyst) has rerevealed an apatite structure in this collophane; on the basis of a chemical analysis, it is believed to be carbonate-fluor-hydroxyl-apatite (sic).

Collophane has an affinity for fluorite, analcite, and zeolites, in the cement, where it forms fine-grained aggregates with them.

Calcite is widely distributed in Middle Devonian and Upper Carboniferous rocks; it has been identified in two generations: 1) sedimentogenic stage; 2) diagenetic stage. Calcite of the first generation is pelitomorphic, commonly finely dispersed in collophane and cryptocrystalline zeolites, and is associated with fine-grained fluorite. In its diagenetic stage, the cement calcite was recrystallized as coarse-grained aggregates locally forming a poikolitic structure.

Dolomite is rare, occurring in rhombohedral crystals in the cement of Lower Carboniferous rocks.

Sulfur compounds are represented by pyrite, chalcopyrite, galena, and sphalerite; they are rather common but occur in small amounts.

III. Distribution of Zeolites and Analcite in Sedimentary Formations of Various Ages

Authigenic analcite and zeolites, both as rock-forming and auxiliary minerals, are known from sedimentary rocks of various ages, from Middle Devonian to Recent. In the Soviet Union, zeolites are most common in deposits on the Russian platform.

Below is a description of the conditions of occurrence of analcite and zeolites in sedimentary rocks.

The Paleozoic. At the present time, analcite and zeolites are unknown from the lower Paleozoic; from the middle Paleozoic, they have been found only in Middle Devonian and Lower Carboniferous sediments of the Tuva Autonomous Oblast'.

Discovered by N. V. Rengarten [17] in Upper Permian deposits from the Kirovsk Oblast on the Russian platform were sandstones with authigenic analcite widely distributed in the cement and associated exclusively with polymictic varieties. A. M. Boldyreva [3] discovered analcite of that age in pelitomorphic limestone, marl, polymictic calcareous sandstone and siltstone of the Chkalovsk and Aktyubinsk Oblasts.

The accumulation of analcite- and zeolite-bearing rocks in the Paleozoic proceeded in epicontinental marine basins (Russian platform) as well as in relict basins at an early stage of their salinity (Lower Carboniferous of Tuva) and in ancient alluvial deposits (Middle Devonian of Tuva).

Analcite and zeolites in Paleozoic deposits occur chiefly in the calcareous cement of terrigenous sequences (less commonly with limestone); this suggests an alkaline medium for its sedimentogenesis and early diagenesis.

Pyroclastic material is almost always present in Paleozoic analcite and zeolite-bearing rocks. Occasionally, it constitutes an important source of original components in the formation of zeolites (Lower Carboniferous of Tuva).

The Mesozoic. Mesozoic analcite- and zeolite-bearing rocks are known from many localities of the Soviet Union and from the U. S. Analcite in rock pores has been observed by M. F. Kolbin and others [13] in the Bogdinsk (Triassic) rocks of Bol'shoye Bogdo Mountain.

Analcite was noted by W. D. Keller [29, 30] in limestone and shale of the Chinle (Utah) and Chugwater (Wyoming) formations, in addition to an essentially analcitic bed (7.5 cm thick) consisting of abundant fine analcite crystals and a reddish-brown cement. Analcitic beds in those areas are marker horizons.

Analcite and laumontite were identified by N. V. Rengarten [19] from Lower Jurassic redbeds of the North Caucasus, where they occur in the cement of ferruginous nodular and sandy rocks.

In the Ural-Emba region, P. P. Avdusin [1] identified the analcite present as the decomposition product of volcanic material in the cement of continental-lagunal $J_1 + J_{2+3}$ sandstones (Dossor formation). As described by G. S. Dzotsenizde and N. I. Skhirtladze [10, 11] from coal measures (Bathonian stage) of the Kutais-Gelat area of Georgia, analcite locally constitutes 80% of the total rock. Here, analcite-bearing pelitic, silty, and fine-grained psammitic rocks alternate with coarse-grained analcite-free sandstone.

G. I. Bushinskiy [6] described mordenite from Upper Jurassic deposits of the Moscow and Volga regions. Analcite and mordenite were also described from Cretaceous deposits of the Russian Platform, the southwestern part of the Ural-Emba region, along the east slope of the Urals, and in western Kazakhstan. Zeolites have been described from Lower Cretaceous sandstone, siltstone, and shale of the Penza Oblast' by V. S. Vasil'yev, M. F. Kolbin, and V. N. Krasnova [8]; these zeolites (not specified) cement clastic material and are enclosed in hollow radiolaria and sponge spicules. The zeolite-bearing rocks are 50 to 80 m thick. G. I. Bushinskiy [6] notes the wide distribution of mordenite in Cretaceous deposits of the Russian platform. V. S. Vail'yev described [7] mordenite from Mesozoic-Cenozoic deposits of the lower Volga and western Kazakhstan. Mordenite was noted in cherty sandstone and phospherite from upper horizons of the Volga Albian.

Mordenite also was described by N. V. Rengarten [18] from Middle Cretaceous (Danian) sandy deposits of the east Uralian slope.

P. P. Avdusin [1] mentions authigenic analcite from Middle Cretaceous shale, marl, limestone, and blackboard chalk from the southwestern part of the Ural-Emba region.

The Cenozoic. Zeolites are no less common in Tertiary deposits than in Cretaceous beds. Here, too, they are usually associated with the Russian Platform (Donbas, Volga region), the east slope of the Urals, the Emba region, and the Caucasus (Georgia, western Azerbaydzhan, and western Apsheron). In addition, zeolite-bearing Tertiary rocks are known from the U. S. (Utah, Colorado, Wyoming, California).

V. S. Vasil'yev [7] described mordenite from Eocene deposits of the Emba region and from the lower Volga Oligocene.

In cooperation with other authors, he also described zeolites, not identified more precisely, from Paleogene deposits of the Penza Oblast consisting of claystone, siltstone, and sandstone; most zeolites here are associated with siltstone.

I. A. Shamray [23] noted that mordenite is widely distributed in Paleogene marine deposits in the southern part of the Russian platform along a belt from the northeast Donbas to Lower Volga; it locally forms essentially zeolitic rocks termed "zeolitite" by Shamray. These zeolitites are traceable in thick beds in the Khar'kov and Buchak-Tsaritsin deposits; according to I. A. Shamray, they are chemogenic sedimentary formations.

On the east Uralian slope, mordenite is distributed regionally in Danian and Paleogene marine deposits [18, 20]. In Georgia and western Azerbaydzhan, analcite and mordenite are common in polymictic sandstone and siltstone [12] where they fill the pore space and

locally replace fragments of extrusive rocks and feldspars.

On the Apsheron Peninsula, analcite occurs with volcanic glass in sands of the productive interval (Pliocene). An essentially analcitic rock from the Eocene Green River formation (Utah, Colorado, and Wyoming) has been described by W. Bradley [24, 25]. M. N. Bramlette, and E. Posnjak [26] described zeolite-clinoptilolite (-heulandite) from pyroclastic rocks of California. Authigenic heulandite from the cement of Upper Miocene sandstones of California has been described by C. M. Gilbert and M. C. Andrews [28].

An analcitic sedimentary rock was described by C. S. Rose [32] from Quaternary lacustrine deposits of Arizona. Like the rock described by W. Bradley [24, 25], it consists exclusively of fine analcite crystals.

According to the Challenger expedition data [31], Recent deep Pacific red clay and radiolaria and globigerina oozes carry phillipsite and locally analcite, both associated with Recent volcanic material and its alteration products.

Thus, detailed mineralogic studies of sedimentary sections have revealed a broad regional distribution of analcite and zeolites as rock-forming minerals.

IV. On the Origin of Analcite and Zeolites in Sedimentary Rocks of Tuva.

We shall now consider certain conditions favorable for the accumulation of analcite and zeolites in sedimentary rocks.

1. Authigenic analcite and zeolites of Tuva are associated with psammitic rocks. Judging from published data, they occur in rocks of diverse grain size, from pelites to psammites. However, they are most common in psammitic rocks with a comparatively high porosity which consequently preserve the moisture in their pores for a longer time.

2. Clastic material of the analcite- and zeolite-bearing rocks usually is polymictic. The presence of ash replaced to a considerable extent by analcite is typical of Middle Carboniferous deposits of Tuva. A number of areas are known from the literature where the formation of zeolites is apparently unrelated to pyroclastic material (Russian Platform, etc.) but is associated with chemogenic formations, limestone [3, 13], blackboard chalk [16], and siliceous marl and chert [7].

3. Associates of analcite and zeolites are calcite, dolomite, phosphorite, fluorite, chlorite, glauconite, chalcedony, opal, hydroxides of iron and manganese, sulfides (pyrite, sphalerite, galena, etc.), clay minerals, and not uncommonly carbonaceous matter. Three paragenetic associations of authigenic minerals have been identified:

1) analcite or laumontite + calcite ± dolomite ± siderite ± sulfur compounds ± barite ± iron hydroxides ± opal, chalcedony ± glauconite ± carbonaceous substance ± minerals [1, 3, 5, 8, 10, 11, 12, 18, 19, 23, 24, 25, 30];

2) analcite, laumontite or mordenite + calcite + phosphates ± fluorite ± carbonaceous matter ± iron oxides ± opal and chalcedony ± clay minerals ± sulfur compounds [4, 6, 7, 31];

3) analcite or mordenite ± glauconite ± opal ± chlorite ± clay minerals ± manganese oxides [18, 32].

The first paragenetic association is the most common.

It is known that authigenic minerals are the indexes of geochemical facies in sedimentation; thus, it has been established that calcite is indicative of an alkaline medium with pH up to 9.5, with pH>9 for phosphorite. These two minerals are the principal syngenetic companions of analcite and zeolites; it may be assumed therefore that their formation is promoted by an alkaline medium.

4. Essentially analcitic and zeolitic sedimentary rocks of Tuva belong to a Middle Devonian — Lower Carboniferous complex corresponding to late and terminal stages of development of a mobile belt (after Yu. A. Bilibin, for the Tuva area [2]). Terminal stages in the development of a mobile belt are similar to platform conditions. According to published data, the formation of analcite and zeolites is of a regional nature in many areas, being related to platform deposits of epicontinental marine basins, lagoons, and lakes of different geologic periods. Herein apparently lies the community of formation conditions (broadly speaking) for authigenic zeolites and analcite of sedimentary rocks in most areas of their occurrence.

5. The presence of analcite in unconsolidated sands of the Apsheron productive interval [16]; of mordenite in Danian and Paleogene sands of the east Uralian slope [20]; and of phillipsite and analcite in recent deep clays and oozes of the Pacific [31], all point to a direct formation of authigenic zeolites and analcite in sediments during an early diagenetic stage (in the N. M. Strakhov concept of the beginning of diagenesis [21]). Analcite nodules (Mt. Ak-Tag) with zoned peripheral segments of alternating fluoritic and analcitic shells described above also suggest their early diagenetic origin.

The texture of analcitic rocks with analcite

porphyroblasts suggests that clots of aluminosilicate gel were formed in undisturbed-bottom conditions; this gel subsequently coagulated and crystallized.

Alumina and silica are usually present in waters of rivers, lakes, and seas; at times their content reaches magnitudes necessary for the formation of authigenic aluminosilicates, i. e., analcite and zeolites. As noted by F. V. Chukhrov, SiO_2 and Al_2O_3 occur in natural waters in both true and colloidal solutions, where "the relationship between the true solutions of aluminum and silicon compounds may lead to the formation of colloidal sediments; however, the colloidal sediments will always have a definite composition corresponding to their stoichiometric relationship" ([22], page 64). Consequently, either form of silica and alumina in solutions may lead to the formation of aluminosilicate gels.

Thus, the formation of zeolites in the Tuva sedimentary rocks was preceded by an accumulation of aluminosilicate gels in bottom sediments of basins, where sedimentary analcite and zeolites passed through a colloidal stage. This is especially well illustrated in essentially analcitic rocks.

SUMMARY

1. Analcite and laumontite are widely distributed in sedimentary rocks of Tuva regionally; they constitute rock-forming minerals important for correlation purposes.

2. Essentially analcitic Lower Carboniferous rocks of Tuva represent facies of drying-out relict basins, possibly lakes or lagoons of a somewhat higher salinity.

3. It has been established that the development of analcite in Lower Carboniferous tuff and sandstone is directly related to submarine alteration of ash material. The formation of essentially analcitic rocks of Tuva has no visible relation to pyroclastic material. Nor do the pyroclastics of other regions necessarily constitute source material for their formation.

4. The formation of authigenic analcite and zeolites was preceded by an accumulation of aluminosilicate gels under bottom conditions.

5. Paragenesis of authigenic analcite and zeolites with syngenetic calcite, phosphorites, and other minerals points to an alkaline genetic environment.

6. Authigenic analcite and zeolites are formations typical mostly of platform sediments and of late and terminal stages in the development of a mobile belt.

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ALTERATIONS IN DIKES CUTTING THE KIZILKOL PYRITE ORE DEPOSIT (NORTHERN CAUCASUS)[1,2]

by

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This paper describes the alteration in dikes of granodiorite porphyry and diabase, in contact with a pyrite ore body. The author believes that these alterations are post-ore in age.

An analysis of physiochemical conditions of bleaching of these dikes shows that it took place in an environment different from that during ore accumulation.

* * * * *

Works of A.N. Zavaritskiy [3], A.V. Pek [7], V.P. Loginov [6], Ye.K. Lazarenko [5], and others [2], on pyrite ore deposits in the middle Urals, contain a comprehensive study of the role of vein ores in the formation of such deposits. The very fact that almost every one of these authors, in studying the same formations, has voiced a different opinion on the relationship between dikes and mineralization shows the difficulty of determining this relationship and above all the relative age of the two.

The urgency for the study of these problems has hardly diminished since the lively discussion on the origin of pyrite, as witness the recent paper by C.L. Knight [9] and the discussion it elicited. He believes that almost all sulfide ore deposits are syngenetic with the sedimentary and volcanic rocks enclosing them. The critical facts on this problem may be provided by the relationship between dikes and mineralization. However, C.L. Knight unfortunately ignores the many works of Soviet geologists demonstrating the presence of pre-ore vein rocks in many sulfide ore deposits, which is in opposition to this thesis.

A standard and reliable criterion in determining the age relationship between dikes and mineralization is the presence of absence of contact alteration in the dike rock. However, such alter-
ation in the dike rock. However, such alterations in dikes in or near an ore deposit may receive an erroneous interpretation.

In his study of the Kizilkol copper-pyrite deposits, this author came to the conclusion that dikes in the ore body had undergone a change much later than the mineralization, and are therefore post-ore. Using the premises of A.G. Betekhtin's theory of hydrothermal processes [1], we shall attempt to explain the physiochemical aspect of these alterations.

GENERAL INFORMATION ON THE ORE DEPOSIT

The Kizilkol deposit is associated with the so-called Peredovoy (Front) Range zone of the North Caucasus and occurs in volcanic rocks, presumably Lower Carboniferous.

The lateral rocks dip steeply but have undergone a comparatively slight regional metamorphism. They are massive to spherical lavas, mostly basic to intermediate in composition. Coarse pyroclastics as well as extrusives of a quartz porphyrite type are prominent in the deposit area. The pyrite ore body, lenticular and broken up by faults into several blocks, is surrounded by an inconsistent, locally missing, halo of pyritized quartz-sericite rocks which may be termed secondary quartzite.

The mineral composition of this ore deposit is comparatively simple. Pyrite is prominent, accompanied with small amounts of chalcopyrite,

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[1] Ob izmeneniyakh dayek, sekushchikh kolchedannuyu zalezh' Kizilkol'skogo (Severnyy Kavkaz).

[2] This paper raises the question of criteria for determining the age relationship between pyrite mineralization and related dikes. The author's observations, however interesting, do not definintely prove a younger age of these dikes, as compared with the mineralization. For the practical purposes of geologists and prospectors, this problem should be given special attention, namely by applying the methods of isotope geology. Russian Editorial Board.

sphalerite, and barite, with bornite and covellite present in the oxidation zone. Comparatively widely distributed in uneven-grained, essentially pyritic ores are collomorphic precipitates of iron disulfide in cryptocrystalline massive to concentrically zoned bodies.

Exploration under the direction of Ye. A. Kryukov has revealed the presence of about ten dikes of extrusive rocks dipping, like the ore body, almost vertically. Prominent among them are granodiorite porphyries and subordinate diabase. Given below is a description of the alterations in these dikes cutting the ore body.

CONTACT ALTERATIONS IN GRANODIORITE PORPHYRY

One of these dikes, about 2 m thick, was traced in mine workings both in pyrite and the enclosing porphyrite (Figure 1). The latter does not show any substantial alteration at the contact with the ore body. Externally, the aspect of the granodiorite porphyry changes radically as the dike passes from porphyrite to pyrite. Instead of a light-green very tough rock, it becomes white and considerably weaker, still preserving large relicts of unaltered granodiorite porphyry. The extent and nature of alteration is apparent in comparing the petrographic and chemical composition of corresponding segments of the dike. In its unaltered portion, granodiorite porphyry in porphyrite consists of individual coarse tablets of albite in a finely-crystalline aggregate of quartz, albite, and chlorite. Also present in this aggregate are prismatic pseudomorphs of chlorite and carbonate.

Evenly dispersed throughout the rock is a large number (about 3% of total volume) of euhedral magnetite grains and rare crystals of zircon, and apatite, with aggregates of leucoxene. Some magnetite crystals have been partially replaced by a dull-white aggregate, spongy in reflected light and strongly diffusive. The importance of chlorite (about 10% of total volume), carbonate, and albite in the mineralogic makeup of the ores apparently points to the participation of granodiorite porphyry in that regional greenstone metamorphism which has affected the volcanic section of this deposit.

The chemical composition of the dike is closest to that of granodiorite, according to R. Daly (see Table 2), and is marked by a high iron and sodium content.

In the pyrite body, this dike maintains its thickness (about 2 m) but is less uniform in constitution. Occurring in its middle are isolated flat blocks, 0.5 to 0.75 m wide, of a drab-green, virtually unaltered granodiorite porphyry. These blocks are enclosed in a bleached rock, without any sharp contact between the two. In its mineral composition and structure, the rock in these blocks is little different from the above-described granodiorite porphyry, except for the characteristic lack of magnetite and other ore minerals. At the same time, granodiorite porphyry in the blocks contains more of the cubic pseudomorphs filled with a dull-white spongy aggregate of indefinite composition. Inasmuch as similar aggregates, as mentioned previously, partially replace magnetite in the unaltered portion of the dike, it appears that these blocks, too, originally carried a considerable amount of magnetite.

There is a zone of gradual transition, 5 to 10 m wide at the contact of the blocks and the bleached rock.

The bleached rock preserves structural features of granodiorite porphyry but has a different mineral composition. Its main constituents are quartz and sericite; albite is preserved mostly in coarse tablets, while chlorite, carbonate, and magnetite are altogether missing.

Dike segments in direct contact with pyrite are schistose and consist on the whole of two minerals, quartz and sericite. The albite tablets here are almost fully quartzitic. Characteristically, pseudomorphs of the dull-white spongy substance on magnetite are also present in bleached rock at the contact.

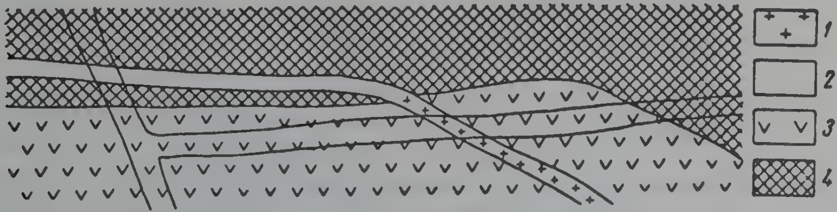

FIGURE 1. Plan of a segment of the ore deposit.

1 - granodiorite porphyry; 2 - bleached granodiorite porphyry; 3 - porphyrite 4 - pyrite.

Table 1

Mineral composition of the granodiorite porphyry

Unaltered	At the pyrite contact
Albite	Quartz
Quartz	Sericite
Fluorite	Albite (relicts)
Carbonate	—
Magnetite	—

Alteration in the mineral composition of the dike is illustrated in Table 1. This table shows that the geochemical process of dike alteration proceeded in the direction of leaching of calcium, iron, magnesium, and sodium. The quantitative and qualitative aspects of these alterations are presented in Table 2 where the chemical composition of unaltered granodiorite porphyry [2] is compared with that for the dike at contact with pyrite [4] and that of slightly altered blocks in the middle of the dike [3].

The table shows that leaching of these elements is compensated for by an increase in the content of silica and potassium. The amount of titanium and aluminum remains more or less constant.

In explaining these phenomena, we take advantage of the well-known concepts of A. G. Betekhtin [1] on the formation of many sulfide ore deposits where hydrothermal solutions are slightly alkaline. Under these conditions of high sulfur activity, iron of the country rocks passes to the ferrous state in forming sulfides. In granodiorite porphyry, magnetite had to be replaced by pryite. Moreover, because of the high iron concentration in solutions during the accumulation of a pyrite deposit, the effect of these solutions on the country rocks would be expressed in additional deposition of pyrite in them. According to J. Murrey and E. Ingersol (as cited in [1]), the stable mineral of an alkaline environment is chlorite, very typical of pyritized rocks.

The observed alterations in granodiorite porphyry, characterized by the silicification and leaching of many other components, first of all

Table 2
Chemical composition of the granodiorite porphyry dike in pyrite [3, 4] and porphyrite [2]

Oxides	1	2	3	4
SiO_2	65.01	64.12	67.28	79.66
TiO_2	0.57	0.45	0.53	0.53
Al_2O_3	15.94	13.34	15.94	12.09
Fe_2O_3	1.74	3.12	0.30	0.19
FeO	2.65	3.66	3.95	0.72
MnO	0.07	0.13		
MgO	1.91	0.90	1.12	0.32
CaO	4.42	3.78	1.79	0.44
Na_2O	3.70	5.34	3.54	2.68
K_2O	2.75	0.98	1.77	1.80
P_2O_5	0.20	0.10	0.15	0.05
CO_2	—	—	1.33	0.05
H_2O^+	1.04	—	1.81	1.64
H_2O^-	—	—	0.48	0.81
SO_3	—	0.20	—	—
$S_{pyr.}$	—	—	0.18	0.11
Losses in heating	—	3.90	—	—
Total	—	100.02	100.17	101.04

1. Granodiorite, after R. Daly [4].
2. Granodiorite from unaltered length of dike (P. P. Ovcharenko, Analyst, N. P. I. Laboratory).
3. Granodiorite from relicts in altered length of dike (after Yu. S. Borodayev).
4. Altered granodiorite (after Yu. S. Borodayev).

iron, could occur only in an environment quite different from that of the mineralization process and obviously subsequent to it.

The recognition of the post-ore nature of the dike alteration in an ore body precludes the use of this criterion in determining the relative age of the dike and the pyrite. Some indirect criteria may achieve that. For instance, even a comparatively thin dike still contains large segments differing but little from its unaltered portion. This fact may suggest a post-ore age because otherwise the probability of preservation of large unaltered blocks in a comparatively thin dike would be very slight. More definite evidence of the comparative age of granodiorite porphyry dikes was obtained from the study of an upper prospecting level.

We shall describe briefly a dike exhibiting the most clean-cut relationship with the ore body. It is 10 to 80 cm thick, with the structure at one cross-section as shown in Figure 2. Prominent in it is massive, homogeneous light-green granodiorite porphyry with altered narrow bands along its hanging-wall contact with the ore body.

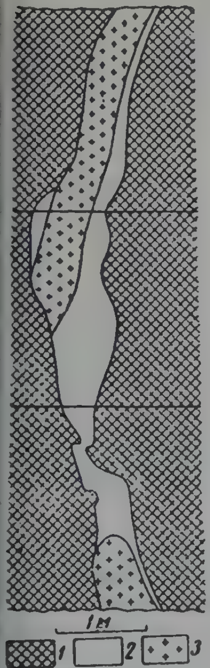

FIGURE 2. Granodiorite porphyry dike in pyrite (evolute).

1 - pyrite; 2 - bleached granodiorite porphyry; 3 - unaltered granodiorite porphyry.

This light-green granodiorite porphyry consists of albite (70%), chlorite (15%), quartz (7%), carbonate (7%), leucoxene, and quasi-transparent spongy aggregates of indefinite composition. In the floor, this rock is in direct and sharp contact with the ore body. The near-contact porphyry has not been altered, although it contains, as does the middle of the dike, thin veins of carbonate and barite.

Bleached segments of the dike are represented by a strongly silicified rock consisting of radial bodies of chalcedony and irregular accumulations of opal with small pockets of hydromica and resorbed albite grains. In addition, this silicified granodiorite porphyry carries a small amount of barite and fine grains of pyrite.

A study of this dike leads to the following conclusions:

1. Silicification of the granodiorite porphyry obviously occurred after the sulfide mineralization, because light-green non-silicified porphyry generally is in direct contact with pyrite.

2. The dike was intruded most likely after the pyrite ore had been formed, because the mineral composition of its light-green segments corresponds to that of granodiorite porphyry of the extrusive rocks. Alterations within the dike (chloritization, albitization, carbonatization) are related to regional greenstone metamorphism of the enclosing volcanic sequence. Consequently, the mineral composition of the dike cannot be a criterion of ore contact metamorphism. The absence of pyritization which is widely developed in extrusives at the contact with the pyrite body suggests that the dike is younger than the sulfide mineralization.

The relative age of some dikes and ore can be determined by mineragraphic study of the ore. A study of a fairly large number of polished sections failed to reveal any substantial difference between ores in contact with the dike and away from it. More specifically, collomorphic formations do not disappear in the contact ores, although they do become less numerous where the ore is coarser-grained. Its mineral content remains unchanged, with pyrite and chalcopyrite present as everywhere else. This, however does not contradict our conclusion. R. Sales and S. Meyer [11] have shown that the contact effect of a rhyolite dike on pyrite-chalcocite ore is expressed in the formation of chalcopyrite-bornite-pyrite ore. Those authors have also shown experimentally that ferrous sulfides of copper are quite stable at high temperatures.

3. The dike carries barite veins. This sulfate mineralization is apparently connected with leaching of the granodiorite porphyry.

ORE CONTACT ALTERATIONS IN VEIN DIABASE

The study of a diabase dike opened in the south contact zone of the ore body reveals even better the post-ore nature of hydrothermal alterations in vein rocks of this deposit.

In the south contact zone, the monomineral ore is pyrite. Contacting it are strongly

pyritized secondary quartzites consisting of 60 to 70% quartz, 10 to 20% sericite, and 15 to 20% pyrite. The diabase dike is located at the pyrite-secondary quartzite contact. Its structure in a cross-section is illustrated in Figure 3.

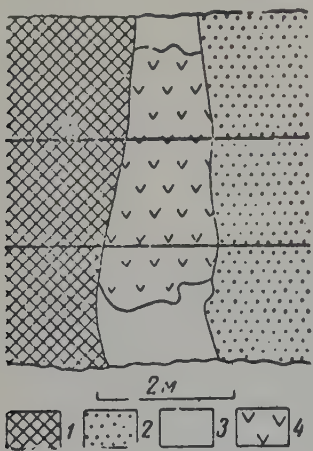

FIGURE 3. Diabase dike at contact of pyrite and secondary quartzite (evolute).

1 - pyrite; 2 - secondary quartzite; 3 - bleached diabase; 4 - unaltered diabase.

The upper part of this dike is represented by a dark-green, homogeneous, massive diabase; the lower part, by a fractured, perfectly white siliceous rock. The boundary between them is an uneven, fairly sharp line running across the dike. Dark-green diabase of the upper part contacts pyrite and secondary quartzite, without any bleaching. This alone indicates a time difference in the sulfide mineralization and the bleaching of diabase.

Morphologically, the dark-green diabase segment most likely is a relict block in a bleached dike. Under the microscope, this rock shows a relict diabase texture of elongated prisms of carbonatized albite in chlorite aggregates. The amounts of chlorite and albite are about the same (40 to 45%).

Carbonate is prominent (about 10%), with quartz and magnetite also present. The chemical composition of this diabase is given in Table 3.

A comparison of data in the table with the composition of diabase after R. Daly [4] shows that our diabase has somewhat less silica and calcium and more aluminum.

Upon bleaching, the diabase is altered to an essentially siliceous rock consisting of quartz, chalcedony, and opal, with a small addition of

Table 3

Chemical composition of diabase dike in pyrite

Oxides	1	2	3
SiO_2	50.48	45.66	87.52
TiO_2	1.45	0.55	0.70
Al_2O_3	15.34	18.74	3.41
Fe_2O_3	3.84	3.57	1.96
FeO	7.78	8.65	0.39
MnO	0.20	0.24	
MgO	5.79	3.64	0.28
CaO	8.94	5.35	Trace
Na_2O	3.07	3.78	1.07
K_2O	0.97	0.58	0.67
P_2O_5	0.25	0.15	
H_2O	1.89	—	—
SO_3	—	0.75	1.06
Losses in heating	—	8.81	3.24
Total		100.46	100.30

1. Intermediate diabase of R. Daly [4].
2. Unaltered segment of diabase dike.
3. Altered diabase (P. P. Ovcharenko, Analyst; N.P.I. Laboratory).

hydromica. Its chemical composition is marked by an extremely high silica content (87.5%). Some of the features of silicification of dark-green diabase are apparent at its boundary with the bleached rock. At this boundary, the diabase has been partially altered: its plagioclase is almost intact while chlorite has been fully silicified and the carbonate leached out. This zone is marked by the presence of a small amount of fine cubes and penta-dodecahedrons of pyrite. In the more strongly silicified diabase, both pyrite and magnetite are missing because of the almost complete leaching of iron.

Table 3 shows that other principal components, with the exception of titanium and potassium, also were subject to intensive leaching. The process of diabase bleaching in pyrite is of the same nature as that described for the granodiorite porphyry dikes.

In order to determine the relative age of diabase and the pyrite mineralization, we shall consider the structure of dike contacts.

The contact of both segments of the dike (dark green and bleached) with the ore body is quite sharp but uneven. As mentioned before, pyrite ore contact is the dike in the north; in the south, secondary quartzites. In either case, the dark-green diabase is not different, macroscopically, from that in the middle of the dike. Nor does a microscopic study reveal any differences.

Immediately adjacent to the contacts is diabase consisting of thin prisms of albite (50%) and chlorite scales (50%), with a very small amount of magnetite and carbonate.

A question arises as to the time of chloritization and albitization of diabase and, more specifically, whether these processes were syngenetic with mineralization.

An answer is provided by a comparison of the mineral composition of diabase and the adjacent secondary quartzite. The latter is characterized by a relict porphyritic and amygdaloidal texture, having been formed in the alteration of extrusive rocks, including those similar to diabase.

Table 4 shows that diabase and secondary quartzite represent different paragenetic mineral associations formed under different conditions.

Table 4

A comparison of mineral composition (in%)

Diabase	Secondary quartzite
Albite, 45 Chlorite, 45 Carbonate, 7 Magnetite, Apatite } Traces	Quartz, 60 to 75 Sericite, 10 to 25 Pyrite, 10 to 20

If the secondary quartzites had been formed after the dike intrusion, diabase would have been pyritized, silicified, and sericitized; thus the absence of such alternation at the quartzite contact points to a post-ore intrusion of the dike. Its chloritization and albitization possibly were synchronous with autometamorphism.

PHYSIOCHEMICAL CONDITIONS DURING ALTERATION OF VEIN ROCKS

In the study of physiochemical conditions for changes taking place in dikes, advantage can be taken of the results of chemical analyses and of the study of mineral composition of dikes and of the features of their occurrence. Dikes of granodiorite porphyry and diabase show a general trend in the change of their chemical composition, with iron, magnesium, and calcium particularly subject to leaching. At the same time, silica accumulates in the rock. The main controlling factor in this process is the effect of the pyrite body itself, because only the dikes located in it are subject to leaching.

As noted by A. G. Betekhtin [1], bleaching processes in the ore-enclosing host rocks are caused usually by low-temperature acid solutions. The alteration of dikes took place long after the mineralization, when the accumulation of sulfides had ceased. At that time, and because of the cessation of the influx of hydrogen sulfide, the oxidation potential in some pyrite deposits should be considerably higher, due to their shallow depth. A penetration of neutral and even alkaline solutions into a pyrite body, along individual channels such as the dike contacts, would cause oxidation of iron disulfide and consequently an abrupt rise in the sulfate anion concentration. Subsequently, these solutions become acid, and leached metal components out of fracture zones at the contact with the ore body. This change in the solutions was accompanied by deposition of silicon dioxide.

Of special interest is the behavior of iron in the course of this process. In oxidation of sulfides, iron forms very slightly soluble hydroxides, an operation requiring, according to the pyrite oxidation potential [8], a very large amount of oxygen. Consequently, it can be expected that with depth and because of a lower oxidation potential, alteration of pyrite would proceed up to the formation of iron sulfate, leached out by the solutions. The behavior of magnetite corroborates this conclusion. Leaching of magnetite appears to have proceeded parallel to the formation of sulfates of ferrous and ferric iron. Oxygen liberated in this process was expended in decomposition of new bodies of pyrite.

The behavior of iron in the alteration of dikes can be compared to that in the formation of pyrite ore bodies, as described by A. G. Betekhtin [1]. In that instance, leaching of iron as a result of the great influx of hydrogen sulfide should virtually cease because oxygen needed for the soluble sulfate compounds of iron is fully expended in oxidation of some of the hydrogen sulfide.

A brief consideration of the chemistry of dike alterations in an ore body, too, convinces us that, first, these changes are post-ore, and secondly that the dikes are younger than the mineralization, because they do not show any trace of the very typical ore-contact effect of pyritization.

It appears, then, that in determining the origin of pyritic and other sulfide deposits, it is very necessary to differentiate between processes contemporaneous with mineralization and the later ones, affected by the presence of an already formed ore body. The qualitative aspect of these processes is different.

A study of ore-contact alterations without considering their origin may lead to erroneous conclusions. An example is the otherwise

interesting paper by H. McKinstry [10], which presents a summary of analyses for the iron content in lateral rocks of 35 sulfide deposits. The author comes to the conclusion that the iron content in altered rocks remains approximately the same or is reduced compared with the unaltered ones, although there are instances of increased iron content. Inasmuch as his paper omits a geologic analysis of the alterations in any of the localities, one may assume that it lumps together genetically different processes. The fact that leaching of iron from ore-contact rocks takes place in many deposits may be indicative of the wide distribution of processes described in this paper.

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GEOLOGIC CONDITIONS FOR THE FORMATION OF SPHERICAL LAVAS IN THE MIDDLE COURSE OF NIZHNYAYA TUNGUSKA RIVER[1]

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This paper cites data showing that the strict association of the Nizhnyaya Tunguska spherical lavas with the lower level of the lava body has been brought about by the tectonic environment (wavy occurrence of lava flows) and interruptions in extrusive activity. This caused the appearance of comparatively small lacustrine basins where the lava flows acquired their spherical structure.

Inasmuch as deposits of Iceland spar are associated with spherical lavas, the latter may be a positive criterion for exploration.

* * * * *

The lava body, which is the upper component of the Tunguska complex extends over vast areas in the north and central parts of the Tunguska syneclise. It is developed generally north of the Nizhnyaya (Lower) Tunguska, where it makes up the Putoran Plateau. This region is rather inaccessible, with poor outcrops, and therefore very inadquately known.

At the present time, there are several schemes of differentiation for individual areas of the northern part of this lava field (or tuffaceous lava of some authors) into members and formations, based on the grain size and degree of crystallization in the basalt flows (K. G. Akimova [1], V. A. Markovskiy and A. A. Mezhvilka, after B. V. Tkachenko et al [15]). M. N. Godlevskiy [4] proposed a new classification for traprock volcanism in the Norilsk region, which postulates several volcanic cycles in its geologic history. Ye. L. Butakova [3] identifies five formations in the Kotuy and Maymecha basin, the northeastern margin of the Tunguska basin, in addition to normal traprock, they contain alkalic and ultrabasic lavas.

In the southern part of the lava field, geologists of the All-Union Aerogeological Trust (I. F. Belostotskaya, A. A. Boruchinskaya, N. V. Drenov, and others) divided it into three formations (reading upward): the Nidym. Kochechum, and Yambukan. The Nidym formation was studied in most detail. Unlike the other two, it includes, along with normal basalt flows, those carrying spherical lavas. Interest in these lavas is due not only to the fact that most of the Iceland spar deposits in extrusive traprocks are known to be associated with spherical lavas: an insight into the conditions of formation of these peculiar rocks will clarify some peculiar features of volcanic activity at initial stages of formation of a lava sequence.

Papers by V. I. Kudryashova [6] and Ye. Ya. Kiyevlenko [5] contain brief descriptions of morphology and petrography of spherical lavas, along with general considerations on their origin (their formation from a lava cooling off under aqueous conditions). At the same time, there is no mention of such important problems as the occurrence of spherical lavas in a certain definite interval of the lava sequence, the Nidym formation, in this instance; the role of tectonics in the process of their formation; regularities in the distribution of spherical lavas within the Nidym formation, etc. These problems have a direct bearing on the history of development at the central part of the Tunguska basin. Their clarification is the purpose of this paper. The data used have been obtained by the author after many years of study of the Iceland-spar deposits along the south margin of this lava field (middle course of the Nizhnyaya Tunguska, from Kiryamka River in the east to Yambukan River in the west).

GEOLOGIC STRUCTURE OF THE TUFFACEOUS LAVA BODIES IN THE AREA UNDER STUDY

Lava sequences including a large number of almost horizontal basalt flow is interbedded with subordinate tuffaceous layers are involved in the geologic structure of this region. These almost horizontal layers (middle interval of the Tunguska sequence), have been observed only at the bottom of the Nizhnyaya Tunguska valley and at the mouths of its tributaries. They

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[1]Geologicheskiye usloviya formirovaniya sharovykh lav Srednego Techeniya r. Nizhney Tunguski.

unlie the basalt flows and are represented by rapidly alternating fine, clastic ash, and agglomeratic vitroclastic tuff, tuffite, and tuffaceous sandstone. Most students believe that both of these sequences are Lower Triassic [14].

The tuffaceous sequence has a very complex constitution, both because of the great variety of its component rocks and their strong facies inconsistency. Only the predominance of some facies in a section can be positively ascertained. Tuffite and tuffaceous sandstone, are conspicuous in upper intervals with lower intervals represented mostly by agglomeratic tuffs. Ash and pisolitic tuff are comparatively rare; lava flows are present only in upper intervals.

Especially widely developed along the Nizhnyaya Tunguska, between Kochechum and Gonchak Rivers (70 km downstream from Tur settlement), are tuffite and tuffaceous sandstone commonly displaying very fine parallel and cross-stratification with polygonal cracks on bedding planes. East and west of there, the relative and absolute proportion of these rocks in the tuffaceous section decreases. Especially conspicuous is the decrease in the coarser clastic facies, to the west, as first noted by A. P. Lebedev [7]. The visible thickness of this sequence fluctuates within a broad range from 5 to 200 m), depending on the erosional truncation and the uneven surface of the sequence.

In getting acquainted with this area, one is struck with the somewhat unusual aspect of lower parts of the Nizhnyaya Tunguska bank slopes. Instead of flat step-like relief, there are broad terrace-like shelves of tuffaceous rocks, with a surface dip of 20° and steeper. Our structural map of the tuffaceous rock surface along the Nizhnyaya Tunguska stretch with tributaries Poledzhikit, Gancha, and Gonchak, shows that these rocks form here folded structures of a domal type, in nearly horizontal rocks (Figure 1); the troughs are less conspicuous.

The domal uplifts are oval in plan, with their long axis extending 2 to 5 km, and short axis 0. 8 to 1. 5 km, with rocks dipping in the flanks at 30 or to more than 40°. The visible closure of these uplifts is 30 to 60 m. The folds show little connection with one another; they have different trends, and show a smaller closure in their upper horizons (Figure 2). The limbs of larger folds (Khavokipr Cliffs near the mouth of Gonchak River) are marked by flexures accompanied by joints and small domal uplifts of the second order.

In noting large dislocations in tuffaceous rocks at the base of the Khavokiprsk Cliffs, G. G. Moor [8, 9] also believed them to be short steep folds of variable trend. The presence of brachianticlinal and domal folds in Mesozoic deposits (tuffaceous sequence) within the lava

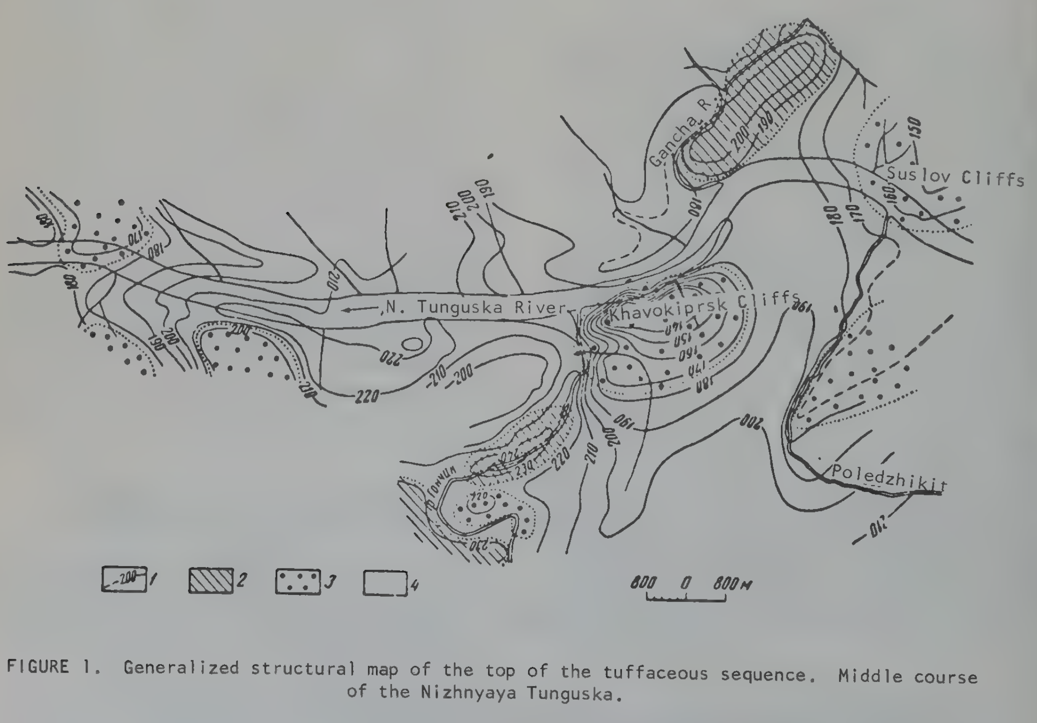

FIGURE 1. Generalized structural map of the top of the tuffaceous sequence. Middle course of the Nizhnyaya Tunguska.

1 - contours on top of tuffaceous sequence; contour interval, 10 m; 2 - domal uplifts; 3 - subsidence areas; 4 - areas of nearly horizontal rocks.

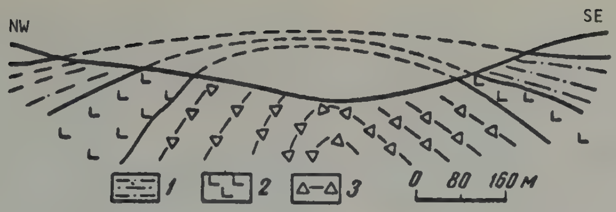

FIGURE 2. Domal uplift, the Gonchak River area, with supratenuous effect.

1 - tuffite; 2 - basalt (flow); 3 - agglomeratic tuff.

field was noted by other students, as well, including I. F. Belostotskaya, A. P. Lebedev [7], and others. At the same time, not enough attention was paid to the causes of these folds which, as will be shown, had played a principal part in the formation of spherical lavas, and this question remained moot. The main difficulty in deciphering these structures is that here we are not dealing with normal sediments but with pyroclastics where the distribution of clastic material depends mostly on the position of the eruptive centers.

The formation of folds appears to have been a protracted process occurring at the same time as deposition of the tuffaceous sequence. This is suggested by the thinning of beds over the folds and by some increase in coarse facies over domes compared with the troughs. The latter phenomenon is not always well expressed. These features of the folds relate them to the so-called supratenuous type ("intermittent" folds of V. V. Belousov) typical of platform provinces and formed in areas of relative warping. According to V. V. Belousov [2] the dimensions of such supratenuous folds, specifically the domal uplifts, may vary from several tens of meters to tens of kilometers.

The origin of such folds is understood differently by different students. Most authors connect them with vertical movements of buried platform basement blocks (N. S. Shatskiy [17]; S. K. Clark [18]; P. Ye. Offman [10]; L. N. Rozanov [12]; N. I. Forsh [16]). The cause of the movements themselves is rather controversial.

On the basis of data extent, we may assume that local folded structures developed in tuffaceous rocks of this area are supratenuous folds whose formation is connected with downwarping of the Kureysk inherited syneclise (the area under study is located on its south margin). According to P. Ye. Offman [11], the latter was formed as a result of non-synchronous subsidence of individual blocks of its ancient basement.

It also may be assumed that similar supratenuous folds are developed north of this area, in a continuation of the Kureysk syneclise. The area of lava flows is confined to that synclease. It has been noted above that domal and brachianticlinal folds are indeed present in tuffaceous rocks of that area.

The formation of folds terminated on the whole toward the beginning of a new stage in the development of this syneclise, at the time when subsidence was replaced by an intensive uplift of the platform, accompanied by abundant flows of lava (the formation of the lava sequence).

The absolute elevation of the top of this tuffaceous sequence remains almost the same throughout the Nizhnyaya Tunguska valley between Kiryamka and Yambukan Rivers (a distance of 200 km); west and east of these points, the top rises, first gently, then at a steeper angle, especially toward the mouth of Viva River. Thus, the surface of the taffaceous sequence in this area, prior to the mass lava flows, was a trough with a flat bottom and isolated small domal uplifts standing on it (supratenuous folds). The boundary of this trough in the present Nizhnyaya Tunguska valley coincides with that drawn for the Kureysk syneclise by P. Ye. Offerman [11].

An angular unconformity is present between folds in tuffaceous rocks and those in lava flows. Because of the local nature of the folding, this unconformity is not present everywhere, being confined to the areas of most intensive folding. In the absence of an unconformity, the boundary between the two sequences is drawn on the predominance of lava flows over tuffaceous rocks.

The lava sequence in this area is divided, as noted above, into three formations made up chiefly of basalt flows and to a smaller extent of tuffaceous intercalations. Only one formation, the Nidym, is considered in detail, because it has the best developed spherical lavas.

The Nidym formation consists of basalt flows of an intermediate thickness (8 to 10 m) interbedded with comparatively consistent thin (0.5 m) tuffaceous sandstone, tuffite, and some conglomerate. Its thickness is 150 to 200 m.

We have studied the constitution of this formation in detail, from its exposures in the Nizhnyaya Tunguska valley, between the mouths of Kiryamka and Yambukan Rivers, a distance of 200 km. The formation is clearly divisible here into two members reflecting different cooling off conditions affecting the lava flows.

The lower, the spar-carrying, member includes two types of lava flows: a) massive basalt (without spherical lava) and b) basalt with spherical lava at the base. A brief description is given below.

a) Massive basalt flows are usually thin, 3 to 4 m and less, and are traceable over short distances (hundreds of meters). These basalts are poikilophitic to poikilophitic-intergertal, less commonly taxitic-ophitic. The upper parts of flows are made up of yellowish- to reddish-brown slag-like varieties with an amygdaloidal texture, intergertal to hyalopilitic under the microscope. The amygdules are mostly calcite with chlorite and bowlingite along the periphery. Less common are palagonite, chlorophane, and zeolites (thomsonite, heulandite, mordenite, etc.). The amygdaloidal zone is 0.5 to 1.0 m thick, on the average. Such zones are locally present at the base of a flow, in which case they are thin, ranging from a few centimeters to 0.5 m (Figure 3-a).

depressions determine the form and size of spherical lava segments; they are approximately lenticular, with their thickness ranging from 0.5 to 10 m, attaining 40 m locally (Nidym River). The areal extent of spherical lavas is equally variable but usually proportional to their thickness. They are either thin lenticular bodies, several tens of meters long, or consistent horizons persisting for several kilometers. The transition from spherical to massive lavas is gradual, usually effected through an amygdaloidal zone of variable thickness.

Thus, the concept of G. G. Moor [9], A. V. Skropyshev [13], and V. I. Kudryashova [6] to the effect that spherical lavas are independent flows has not been corroborated by our detailed study. Such concepts do not fully explain the coincidence of spherical lava "flows" with corresponding depressions; nor do they account for their lenticular form.

Spherical lavas consist of individual spheroids (less commonly spheres) of basalt with an average long axis of 0.7 to 1.0 m (Figure 4). The central parts of these spheroids are made up of tholeiite, rarely poikilophitic-intersertal basalts, with amygdules along the periphery. A thin (1 to 1.5 cm) vitrophyric hyalobasalt outer shell has been observed locally. Each spheroid has a typical radial (less commonly

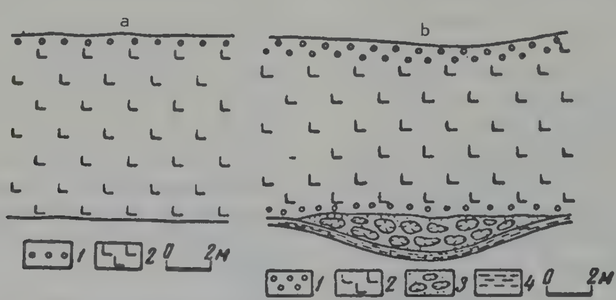

FIGURE 3. Details of the flow structure.

a - dense basalts (without spherical lava); b) basalt flow with spherical lava at base; 1 - amygdaloidal basalt; 2 - basalt; 3 - spherical lava; 4 - tuffaceous sandstone.

b) Basalt flows with spherical lavas at their base are comparatively thick (6 to 45 m) persisting at times over tens of kilometers and constituting marker horizons. Their only difference from the first type of flows is the presence of spherical lavas at their base.

Spherical lavas are not developed along the entire extent of a flow. They are always associated with depressions in the underlying flow or tuffaceous layer (Figure 3-b) and may be missing where the underlying surface is level or bulging. It has been observed that these

concentric) arrangement of joint planes, suggesting an individual cooling surface for each one.

The spheroids are either in close contact or separated by unconsolidated, finely clastic material which we believe to be a disintegration product of formerly thicker, vitreous shells of these spheroids. Their small fragments, represented by hyalobasalt, are fractions of a centimeter to 1.5 cm in size cemented by hydrothermal minerals which form very fine veins, aggregates, and pockets. The most common of

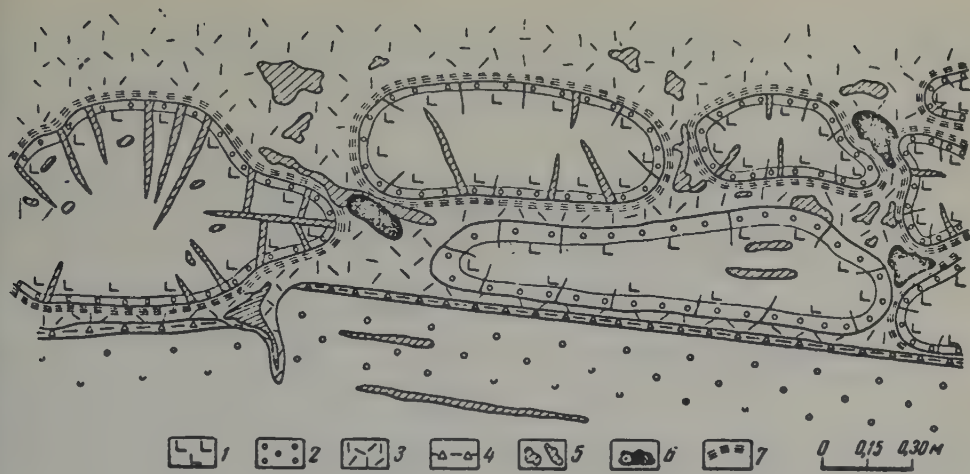

FIGURE 4. Structure of spherical lava. Individual basalt spheroids with finely clastic mineralized material between them.

1 - basalt; 2 - amygdaloidal basalt; 3- finely clastic material; 4- tuffaceous sandstone; 5 - aggregates and veins of hydrothermal minerals; 6 - pockets of calcite and Iceland spar; 7 - broken up vitreous shell of a spheroid.

these minerals are zeolites (chiefly mordenite), chlorites, calcite (including its transparent variety, Iceland spar).

The structure of these spherical lavas is conspicuous because of its lack of uniformity. Present in different lenses as well as within a single lense, is a different quantitative relationship between spheroids themselves and the interstitial segments of finely clastic material; also variable are the dimensions and forms of individual spheroids and the thicknesses of their amygdaloidal shells.

The following three types of spherical lavas can be identified:

1. Spherical lavas of type one make up lenticular units, over 2 km long and 10 m thick and more. They occur only at the base of the lower Nadym horizon (Khavokiprsk Cliffs, near the mouth of Gonchak River, a tributary of the Nidym). They display a fairly definite vertical zonation, expressed in both the distribution of spheroids of different kinds and of minerals cementing the finely clastic material.

Located in lower intervals of these units are the largest spheroidal to pillow-like bodies of lava packed very close to one another. Their long axes are up to 5 m (Figure 5-a). Individual bodies consist of basalt without an amygdaloidal shell but with a well-developed knobby vitreous crust.

Higher up, in the middle interval of a flow, the spheroids are smaller (long diameter up to 0. 7 to 1. 0 m) while the spaces between them, filled with finely clastic material, are larger; here, the spheroids here have an amygdaloidal shell.

In the upper part of a layer, the spheroids are flattened to pillow- or pancake-like bodies without zonation and consisting completely of amygdules. The space between individual bodies contracts and the lavas lose their spherical aspect to become massive amygdaloidal basalt (a narrow transition zone of 0. 5 to 1. 5 m) and then massive basalt.

The minerals cementing the finely clastic material, too, exhibit a definite regularity. Mordenite is present everywhere but is most conspicuous in lower parts of flows where heulandite is also present. Appearing in the middle part are apophyllite, laumontite, and chalcedony. Higher up, they are replaced by analcite, thomsonite, and stilbite. Calcite and Iceland spar usually occur in the middle part of flows; they are present locally in massive amygdules.

2. Spherical lavas of type two are represented by shorter lenses, 4 to 5 m thick and 500 to 600 m long. These lavas are typical of the lower and particularly the middle part of the lower Nidym unit (Gonchak River). They are more uniform throughout the section, being represented by approximately equal spheroids (diameter 0. 5 to 1. 0 m) with a poorly developed amygdaloidal shell and a small amount of interstitial material, so that the spheroids are usually in contact (Figure 5-b). At the base of a lens, individual spheroids are elongated into rollers leveling up the uneven underlying

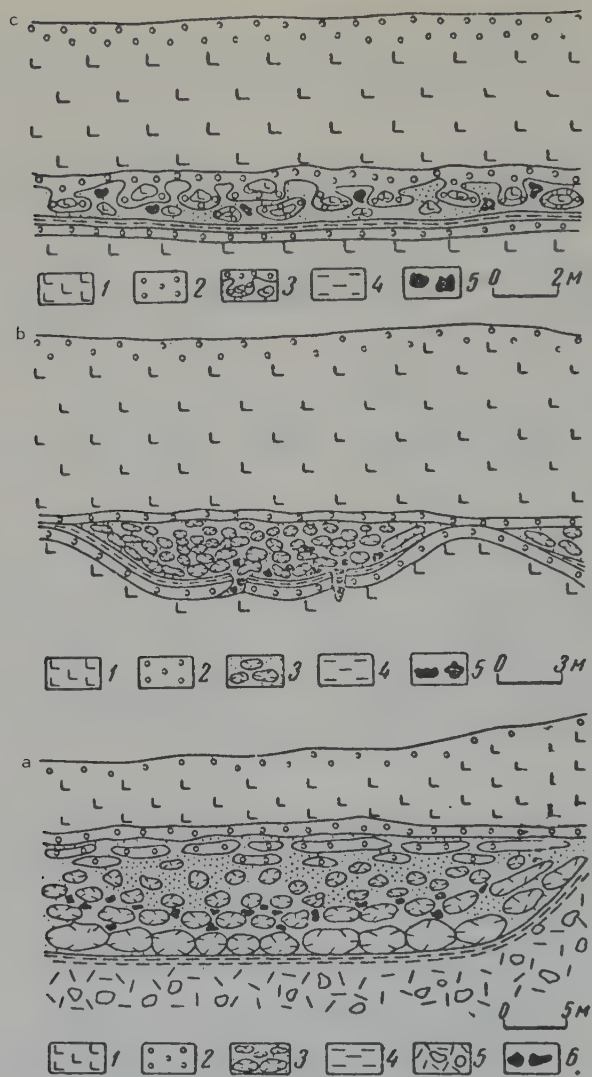

FIGURE 5. Types of spherical lavas and the mode of their occurrence

a) a thick lenticular layer of spherical lava in the Nidym River area: 1 - basalt; 2 - amygdaloidcal basalt; 3 - spherical lava; 4 - tuffaceous sandstone; 5 - pockets of calcite and Iceland spar; b) short lenses of spherical lava, the Gonchak area; 1 - basalt; 2 - amygdaloidcal; 3 - spherical lava; 4 - tuffite; 5 - pockets of calcite and Iceland spar; c) thin flat lenses of spherical lava, Suslov Cliffs, Nizhnyaya Tunguska; symbols the same as in b.

surface. Where this surface is very rough, sharply warped, and therefore contains open fractures, lava may penetrate these fractures as deep as one meter. Spheroids formed in such cracks are smaller (0.2 to 0.4 m) and usually made up of amygdules; interstices between these spheroids are filled with finely clastic material, as usual. Here, chalcedony is the important cementing mineral. The wedging out parts of open fractures are filled with mineralized, finely clastic material free of spheres. Predominating here are chalcedony and calcite, with mordenite occurring in smaller amounts.

It is of interest that Iceland spar crystals are often associated with these peculiar lava intrusions known as the fracture variety os spherical lavas. If present at all, Iceland spar crystals occur in abundance.

3. Spherical lavas of type three are represented by flat thin lenses, from 0.5 to 1.5 - 2.0 m thick and up to 100 m long. They have been observed in the upper part of the lower Indym unit (Suslov Cliffs, the Nizhnyaya Tunguska, and Poledzhikit). Here, the spheroids are mostly not fully separated from the overlying basalts but rather occur in reniform growths (Figure 5-c). They are up to 1.5 m in length and are marked by a well-developed amygdaloidal shell; smaller spheres (up to 0.3 m) are fully amygdaloidal. Spherical lavas of this type carry a large quantity of Iceland spar crystals between the spheroids. Mordenite and heulandite are common among minerals cementing the loose material.

It appears, then, that all three types of spherical lavas occupy a definite position in section of the lower unit. Wherever observed, they are separated by two, three or more standard beds, i.e., free of spherical lavas.

Lava flows of the lower Nidym unit are marked by their gentle folding, duplicating to various degrees the tuffaceous sequence surface relief. Judging from their generally northeasterly dip, their source lay southwest of this area. The average surface slope appears to be only 1 to 2°.

The moving lavas, encountering depressions and highs on their way (local domal folds separated by troughs) preserved to a certain extent the tuffaceous sequence surface relief. This preservation tendency was manifest in all subsequent flows. All flows within the lower units are somewhat upwarped over domal uplifts and downwarped in troughs. Each subsequent flow was less affected by the ancient relief, with only slight undulations noticeable in the uppermost flows of this unit.

Tuffaceous beds, represented by tuffite, sandstone, and conglomerate, always directly underlie the spherical lavas, in depressions of the flow surface underneath them. Microscopic study of tuffaceous rocks shows that rounded hyalobasalt fragments are prominent in the clastic material; these fragments appear to be mostly disintegration products of the upper vitreous crust of lava flows. The cement is mostly calcitic, developed on ash material. The relatively well-rounded material, together with the presence of tuffaceous conglomerate suggests that these tuffaceous rocks were deposited in small fresh-water lacustrine basins with some fluvial deposits.

These data point to intermittent extrusive activity during the initial period of formation of the lava sequence (or, more precisely, of the lower Nidym unit). Lava flows, rapidly following each other, occurred in groups separated by long periods of quiescence when the top of the youngest flow was broken up and carried away to the depressions occupied by basins. A portion of pyroclastic material was brought in as dust from other areas of the Tunguska basin where volcanoes were still active. Each individual volcanic cycle opened with the thickest flow of lava. The different cooling-off conditions of that flow, partly underwater (in depressions); partly on land, determined in their turn the difference in the structure of its segments.

The portion of the flow above water level crystallized as massive and amygdaloidal basalt, while the lower portion, in the water, acquired a spherical structure. This was followed by a rapid succession of smaller lava flows which solidified in sheets of massive basalt free of spherical lavas. A period of erosion followed. The formation time for the lower Nidym unit was marked by three basic volcanic cycles corresponding to the three types of spherical lavas.

The total thickness of groups of flows corresponding to individual volcanic cycles increases upward; its average is 30 to 40 m. The total thickness of the lower Nidym unit is inconsistent, ranging from 80 to 140 m.

The upper Nidym unit differs from the lower in its rhythm and the nature of volcanic activity, as expressed in the ascendency of tuffaceous rocks, in their change in composition, the disappearance of spherical lavas, etc.

This unit consists of basalt flows, in groups of two to four, total thickness 20 to 30 m, separated by tuffaceous rocks, which are ash tuff, tuffite, and subordinate tuffaceous sandstone. They are coarse-grained, locally fine-grained in thinner beds, horizontally stratified, less commonly cross-bedded. Unlike the lower unit, clastic material occurs mostly in not well-rounded fragments of volcanic glass; also less common here are vitrobasalt fragments and isolated grains of quartz and feldspars. The cement is chiefly chloritic, developed on fine ash material. The tuffaceous beds range in thickness from 2-3 to 30 m, becoming thinner upward. The amount of ash tuff, as compared with tuffite and tuffaceous sandstone, increases in the same direction.

Rocks in the upper units are virtually horizontal, showing only a slight northeasterly dip (12 m in one kilometer). This monoclinal dip appears to be due to the general structure of the Kureysk syneclise and the position of the flow source. Ancient relief was no longer affecting the position of rocks. The average total thickness of the upper unit if 100 m.

Above that, there are rocks of the Kochechum and Yambukan formations which we shall consider only briefly.

The Kochechum formation, the thicker of the three, corresponds to the period of the largest flows.

Its boundary with the underlying Nidym formation is drawn on the base of a 80 to 90 m thick flow traceable over tens of kilometers where it crowns the Nizhnyaya Tunguska banks. This flow consists of basalt porphyry with coarse incrustations of plagioclase in a taxitic groundmass; the content of olivine which locally forms glomeroporphyritic aggregates is 20%. Some students describe this flow as an individual unit of basalt porphyrite.

Thick lava sheets alternate with thin terrigenous deposits and members of thinner flows. The formation thickness ranges from 300 to 400 m in the lower Kochechum course to 400 to 800 m and better in northern areas.

The Yambukan formation represents a period of comparatively weak extrusive activity. It is made up of homogeneous sheets of average thickness, interbedded with thin (0. 5 to 1. 5 m) tuffaceous sandstones consistent in thickness over considerable distance. This formation is 80 to 120 m thick.

The Kochechum and Yambukan formations are almost horizontal, with a slight northeasterly dip of 6 to 8 m per kilometer.

SUMMARY

1. The association of spherical lavas with topographic lows in the surface over which the lava flowed, and their constant association with tuffite, tuffaceous sandstone, and tuffaceous conglomerate, all showing evidence of subaqueous origin (fine parallel- to cross-stratification, fine ripple marks on bedding planes, freshwater fossil fauna and flora, etc.), suggest that the spherical lavas themselves were formed in contact with water.

2. The stratigraphic position of the spherical lavas is quite definite: they occur only in the lower unit of the Nidym formation. This is explained by the fact that the flows in this unit, in their progress over relief built upon folded rocks acquired an undulating aspect. Because of an alternation of gentle highs and lows and also because of the intermittent nature of the extrusive activity, numerous small freshwater basins were formed here, between the volcanic cycles. Since each flow leveled off the relief, the position of flows became progressively more horizontal. For this reason, and also because the rhythm of the extrusive activity changed progressively (steady flows changing to sporadic outpourings, then back to flows), basins ceased to exist, and the spherical lavas with them.

3. The data extent suggest that similar geologic conditions prevailed in adjacent areas of the lava field, north of the Nizhnyaya Tunguska. The area of spherical lavas in the Tunguska basin is considerable. Latitudinally, it trends between the Krasnyye Stolby post in the east and Viva River in the west.

4. A characteristic feature of spherical lavas is their strong mineralization (chiefly by zeolites and Iceland spar). We believe that one reason for this was the high content of volatiles, primarily water vapor, in the upper portion of the magmatic hearth, even prior to the outpouring. Consequently, the first lava flows at the base of the lava sequence contained more volatiles than subsequent flows.

Furthermore, the flows of that period were of a pulsation type, i. e., alternating with long quiescent periods. This promoted a new concentration of volatiles in the upper part of the magmatic hearth, to be carried out in the very first flow of a new cycle, usually transformed into a spherical lava flow sheet.

In addition, the intensive mineralization of the lower lava unit, and particularly that of the spherical lavas, is explained by the fact that the quantity of hydrothermal solutions increased considerably because of the assimilation of water from freshwater basins into which the lava was pouring.

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BRIEF COMMUNICATIONS

CROSS-STRATIFICATION IN UPPER JURASSIC LITTORAL MARINE DEPOSITS OF SOUTHEASTERN UZBEKISTAN[1]

by

V.I. Troitskiy

Upper Jurassic deposits in southeastern Uzbekistan consist of limestones 300 to 600 m thick. Most students [1, 2] believe them to be true marine deposits.

We have observed in these limestones, beds of gravel and sandstone traceable for a distance of 15 km along the Surkhantau anticline. They consist of clastic material with a small amount of carbonate cement. Predominant in the clastic is quartz (70 to 75%) with some feldspars 10 to 12%) and occasional fragments of siliceous rocks (3 to 5%). The grains are subrounded to rounded; the cement is carbonate, represented in pores by fine-grained calcite. There are occasional fragments of pelecypod shells.

This unit exhibits cross-stratification, with the cross-bedded series locally as much as 8 to 10 m thick. Such phenomena are little known. To be sure, the works of A.V. Khabakov [5], L.B. Rukhin [4], and others mention that such a thick stratified series may originate under marine littoral conditions. The reason for designating it an independent type is its great thickness and the wide development of ripple marks on its oblique bedding planes.

These oblique layers are rectilinear, parallel to one another, and are several centimeters thick (Figures 1, 2). The bedding planes carry mud-eaters' tracks and symmetrical ripple marks oriented along the strike. The dip of these cross beds is 5 to 15° or more. In a cross-section normal to the dip, they are rectilinear, oriented parallel to the strike,

[1]Kosaya sloistost' v pribrezhno-morskikh otlozheniyakh Verkhney Yury Yugo-Vostochnogo Uzbekistana.

FIGURE 1. Monoclinal cross-stratification in sandstone.

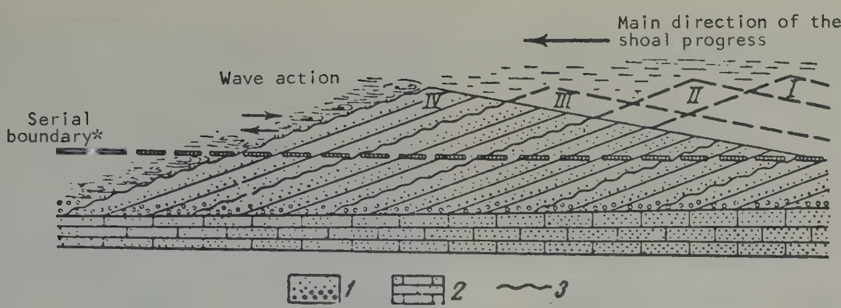

FIGURE 2. The origin of cross-stratification.

1 - sandstone and gravel; 2 - sandy limestone; 3 - ripple marks; I-IV - stages of the bar progress.

*Translator's Note: Boundary between finer and coarser material in a series.

clearly distinguishable, and definitely traceable within the series.

Material in cross beds is usually well sorted, with quartz pebbles and shell fragments present at the base of a series. The serial boundaries commonly extend over considerable distances (up to several kilometers), with the series thickness maintained at 5 to 10 m.

We believe that such stratification originates under the conditions of sedimentation which form submarine bars. Particles in the bottom load, carried by the wave current, are transported up the slope facing the sea [3] and are deposited on the opposite slope to form a cross bed, on which ripple marks are formed by wave action. This explanation, however, is not the only probable one. It is possible, for instance, for cross-bedding to be formed in the action of a tide advancing over a beach. However, in that event, the dip of the cross-stratification should be considerably flatter. In any event, this type of stratification suggests the development of littoral-marine deposits in the Upper Jurassic limestone section of southeastern Uzbekistan.

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RHYTHMIC PHENOMENA IN A COARSE-DISPERSED MEDIUM DURING THE FORMATION OF LIMONITE GEODES IN THE BILIMBAY CRYSTALLINE[2] LIMESTONE

by

G. N. Vertushkov

The banded texture of mineral substances is often determined by diffusive and capillary processes in porous media [5]. Some compounds, on being diffused in such porous media, form rhythmic (banded) textures similar to Liesegang rings [1]. Iron compounds whose hydrolysis

[2]Ritmicheskiye yavleniya v grubodispersnoy srede pri obrazovanii zheod limonita na Bilimbayevskom mestorozhdenii kristallicheskikh izvestnyakov.

produces limonite are examples. A typical periodicity which takes place in such mineral deposits has been observed in brown, metasomatic ferruginous geodes in limestones near Bilimbay settlement, on the west slope of the middle Urals. Here, a quarry in the terrace above the floodplain on the right bank of Chusovaya River has uncovered a karst bed of crystalline limestone in ancient formations of the axial part of the Urals. Red-brown clay which covers the limestone and fills up the karst hollows contains a large number of limonite geodes and crusts. The crusts, to judge from their shape, are fragments of geodes.

Crystalline limestone of this locality have a clean-cut bedding striking meridionally and dipping east at 50°. A system of joints, parallel and across the strike, break up the limestone body into almost rectangular parallelepipeds. Weathered limestone is white, highly porous and hygroscopic, and readily pulverized by a hammer blow. The unaltered limestone is dense, rings when struck, and is grayish white, with a brownish cast. Under the microscope, this rock is granoblastic, with isometric calcite grains about 0.05 mm in diameter. The chemical composition of this limestone is almost exactly calcium carbonate; chemical analyses of averaged samples yield very small amounts of MgO, MnO, SiO_2, Al_2O_3, etc. The average iron oxide content does not exceed 0.2 to 0.3%.

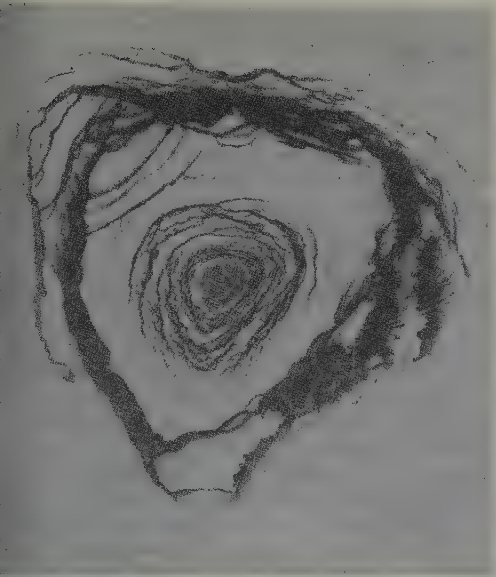
The weathered limestone is saturated with limonite whose content decreases sharply with depth. In dense limestone, limonite is present only in fractures, where its precipitation is quite similar to the Liesegang rings (Figure 1), except that these "rings" are present here in recurring, elliptically concentric crusts 2 to 3 mm thick. These limonite crusts are two or thre three times wider than the intervals between them, and they become thinner going in from the periphery of each ring. These eppliptic "crusts" are wavy, commonly broken, and variable in width. Adjacent "rings" are interconnected, so that it is impossible to trace each individual ring, although its general outline resembles the ideal form for an over-all concentric-rhythmic pattern. Where deposited in fractures, limonite often replaces calcite. The rings are best developed in fractures parallel to the stratification. As a rule, the major axis of elliptic crusts coincides with the dip of the joint plane, and their lower portions are more massive. Similar limonite rings have been observed in fractures of the Karaul'naya Mountain quartzite. It follows, then, that this particular shape of the limonite structures is determined by properties of the iron compounds out of which limonite originates, as well as by the presence of capillary channels where the movement and deposition of mineral substances takes place. Limestone, as the host rock, merely promotes a larger accumulation of

FIGURE 1. Diagrammatic sketch of limonite deposit in a fracture of dense limestone.

limonite (e.g., compared with vein quartz) without affecting its form.

In weathered porous limestone, limonite is distributed as unevenly as in the fractures. Its deposits in a porous rock are spheroidal to ellipsoidal. In such spherical bodies, limonite-saturated limestone layers alternate concentrically with barren layers. The structure of such limonite accumulations is similar to that described for fracture formations (Figure 2). As a rule, the outer limonite layers are merged generally forming a massive crust fully enveloping the entire body. Such formations are typical geodes in limestone. The interior of the geodes contains one or two central concentric limonite accumulations (Figure 3). The structure of one of them (e.g., the geode represented in Figure 2) is illustrated in Figure 4. The form and relationship of limonite bands are distinct in the photograph, along with the well-defined tendency of the mineral substance to gravitate toward the center of the geode, as expressed in convex bends of each layer. This produced the wavy aspect of the shells.

Most geodes developed in limestone approach a triaxial ellipsoid in shape. Its major axis lies in the bedding plane, with the intermediate



IGURE 2. Concentric deposits of limonite in porous limestone.

ight color calcite; dark rings, limonite; reluced 2.25 times.

is perpendicular to it. The position of the inor axis is difficult to determine; it appears run 20 to 40° to the strike.

Concentric deposits of limonite in a homoenous medium should be circular when formed fractures, and spherical in the rock itself. owever, under natural conditions, these forms e distorted due to the effect of gravity, the lack of uniformity in the distribution of pores and cavities, and finally the form of the host block.

As an effect of gravity, rhythmic limonite formations are elongated vertically, and a geode acquires a "top" and a "base." When a limonite crust is consolidated, mostly the base of a "ring" is enlarged in fractures, and its upper part when developed in the rock itself.

In the vicinity of fractures and breaks in the homogeneous limestone, limonite crusts are strongly convex toward the center of a geode. As noted in some specimens, the origin of two central limonite structures in a single geode has been determined by the presence of a fracture in the middle of the host block (Figure 3). Here, the opposite sides of the geode come close together, in the direction of the fracture, and form two independent centers of attraction, symmetrical in relation to the fracture. In porous limestone, the natural boundaries of each concentric limonite deposit are fractures, because it was from them that iron compounds diffused toward the center of a more or less homogeneous block. For this reason, some geodes are reminiscent somewhat of the parallelepiped limestone blocks.

Away from the dense limestones and toward the surface, limonite deposits in fractures gradually disappear to give place to brown iron ore which permeates the rock. The higher the porosity of the limestone, the more massive are the concentric limonite deposits. In the most shattered zone, the upper part of geodes is exposed, with the exposed parts made up of a more compact and massive limonite compared with the lower parts, which are still enclosed in limestone. In the weathering of limestone, most

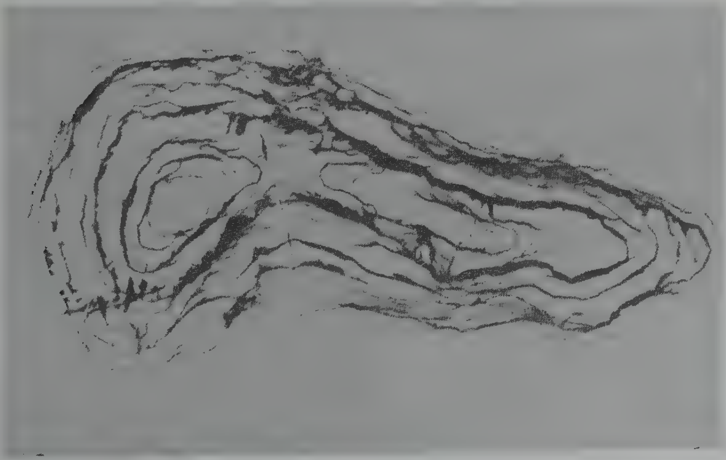


FIGURE 3. Concentric deposits of limonite in limestone, with two central parts developed symmetrically with relation to the fracture.

Light color, limestone; dark, limonite; reduced 2.25 times.

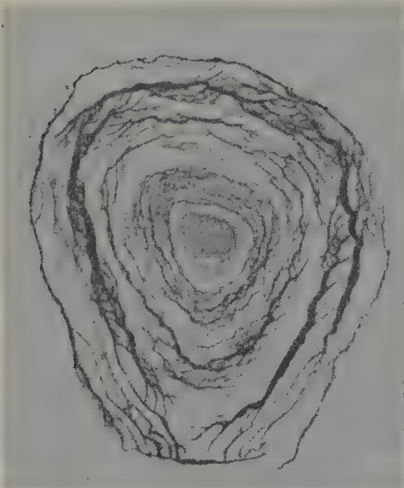


FIGURE 4. Concentric deposits of limonite in porous limestone.

Central part of geode represented in Figure 2. White, limestone; black, limonite; 2.25 × [probably "reduced 2.25 times," in analogy with Figures 2 and 3].

calcite is leached out of the limonite; as a resu[lt] cavities originate within the spherical bodies of limonite. Such "free geodes" continue their development in the limestone eluvium. Their outer parts accumulate limonite contaminated with clay and quartz pebbles, while their interio[rs] are sometimes filled with quartz and clay. Mineral matter is precipitated on the inner walls of hollow geodes, as in miniature caves (Figure 2). A more massive crust of goethite and lepidocrocite is formed on the "ceiling" (often in stalactites), while limonite with clay inclusions is deposited on the "floor."

The formation of geodes in the Bilimbay deposit is related to the weathering of limestone where a dense rock becomes porous. Into this porous body, surface water carried iron compounds and precipitated limonite which metasomatically replaced calcite. The deposition of limonite in these porous limestones proceeded according to the same laws controlling the deposition of gels. As the weathering of the limestone progressed and new batches of iron compounds arrived, the earlier rings were consolidated unevenly, with the outer layers growing more rapidly. As a result, limonite geodes were gradually formed in limestone. As more calcite was dissolved, these geodes eventually became detached from the environment of their origin.

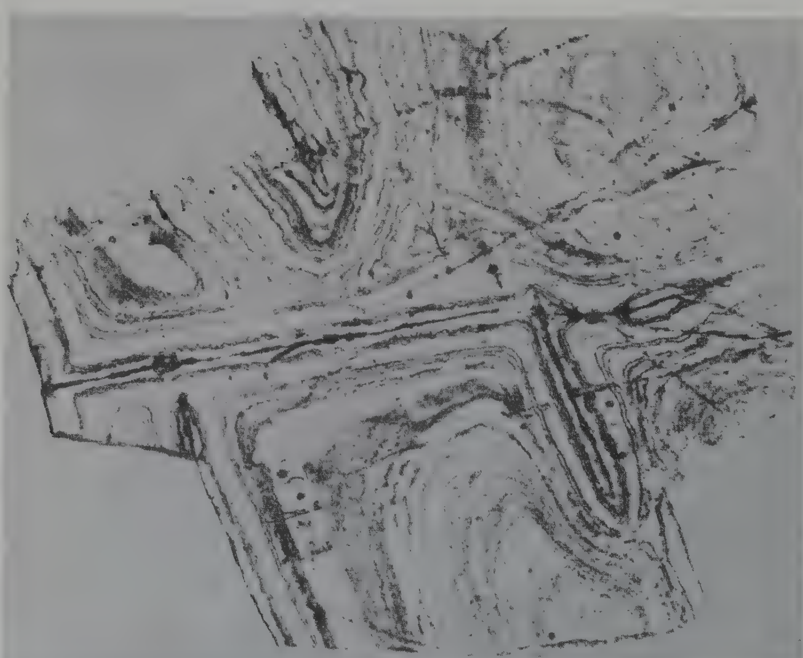


FIGURE 5. Concentric deposits of limonite in quartz-sericite schist.

Light color, quartz-sericite schist; dark, limonite; natural size.

As early as the 1930's, L.V. Pustovalov [4] noted the origin of geodes in sandstone and limestone. Rhythmic deposition of iron hydroxides in sandstone was described by K.P. Yanulov and I.A. Rotosh [6]. We came across an initial stage of geode formation in quartz-sericite schist from Sysert'. Present here is definite evidence of the centripetal progress of iron compounds in the rock body. Our specimen of this Sysert' quartz-sericite schist is broken up by joints (almost vertical to the schistosity) into parallelepipeds. These joints contain massive limonite aggregates (Figure 5) while the rock itself has periodically recurring bands of this mineral. In the vicinity of the joints, the limonite-rich bands duplicate angular form of the blocks.

Away from the joints, sharp angles formed by the limonite bands are less conspicuous; well within the parallelepiped block, limonite bands are annular. Similar rings of limonite were observed by B.V. Chesnokov in the weathering zone of berezite, in the Berezovsk gold deposits, the Urals. This suggests that limonite deposits in rhythmic bands are not due to the chemical composition of a rock but rather to its porosity and the properties of iron compounds from which the limonite is formed.

The banded distribution of brown iron precipitated in a porous medium determines the bizarre grain pattern of some rocks. A good example is the Fominsk marble of the Urals. The intersection of two systems of limonite bands indicates recurrent stages of iron hydroxide deposition in a porous medium whose properties as well as the composition of the solutions changes substantially.

The formation of rhythmic deposits of ferruginous minerals in diffusive and capillary processes in a coarse medium provides a new approach to the origin of zonation in some mineral bodies. Regularly recurring layers in metamorphic rocks, different in composition but corresponding to a single stage of metamorphism, could have been the result of diffusive processes in a rock homogeneous in composition and structure.

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The Sverdlovsk Mining Institute

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LOSSES TO SCIENCE

Sergey Alekseyevich Dobrov died on June 26, 1959. He was doctor of geologic and mineralogic sciences, formerly senior scientific worker of Moscow State University.

Dobrov was born on May 28 (16), 1884, graduated from the Moscow University (Department of Mathematics and Physics) in 1912 and a short time later was appointed assistant to the chair of geology, and then curator of the geological museum of the University managed by academician A. P. Pavlov. In 1914, he began to work on assignments for the Geological Committee, and, became member of its Moscow branch in 1918. During the first years of his scientific activity S. A. Dobrov worked on compilation of geologic soil maps and search for phosphorites in central Russia. For many years he was engaged in phdrogeologic and construction studies, working in particular on the construction of the Moscow Canal.

His paleontologic studies of the fauna of the Upper Cretaceous Inoceramus zone of the Russian platform are of considerable scientific importance. He also carried out many extensive paleontologic determinations which served as the basis for stratigraphic conclusions of a large number of researchers who sent him their collections for processing.

S. A. Dobrov wrote 35 printed works, including chapters in Atlases of Leading Forms and the fourth volume of Geology of the U. S. S. R. In 1952, he was pensioned but continued his paleontologic studies.

Mark Isidorovich Rokhlin, candidate of geologic-mineralogic sciences died on August 13, 1959. He was a member of the Communist Party of the Soviet Union since 1941 and senior scientific worker of the Research Institute of Arctic Geology. He was born on November 22 (9), 1911, graduated from Leningrad Mining Institute in 1932 and started working on exploration for tine and rare metal deposits in Kareliya and Zabaykal'ye. Starting in 1936, he worked for 20 years in the Arctic, mainly in Chukotka studying problems of regional geology, petrography and metollogeny.

M. T. Rokhlin was the author of two books on tin deposits of the northeastern part of the U. S. S. R., and of many printed works and manuscripts. He was awarded the order "Znak Pocheta" (Sign of Esteem) and medals.

Sattar Mirza Ogly Aliyev, candidate of geologic mineralogic sciences died on August 27, 1959. He was a member of the Communist Party of the U. S. S. R. since 1939, and lecturer of the Azerbaydzhan Institute of Oil and Chemistry. Aliyev was born on August 12, 1912, graduated from the Geological Survey Faculty of the Azerbaydzhan Industrial Institute in 1949. For several years he worked as senior geologist of the trust "Aznefterazvedka" and in 1953 switched to pedagogic work and taught the following courses: Prospecting and Surveys of Oil and Gas Deposits, Structural Geology and Geotectonics with Principles of Geomorphology.

S. M. Aliyev participated in the study of the problems of oil occurrence in Miocene, Oligocene and Mesozoic deposits of northeastern Azerbaydzhan and in investigations of the possibility of oil deposits in the area of Kobystan and Nizhnyaya Kura. He also worked on computing reserves of a number of large industrial deposits of oil and gas.

S. M. Aliyev wrote the textbook Structural Geology in the Azerbaydzhan language. He was awarded two medals of the Soviet Union.

On August 29, 1959 doctor of biological sciences, Professor Aleksey Petrovich Bystrov died. He was manager of the paleontologic laboratory of the geology faculty of Leningrad State University. He was born on January 31, (19), 1899, graduated from the Military-Medical Academy in Leningrad in 1926 and then studied comparative anatomy. His main work delt with paleontology of vertebrates: fishes, amphibias, reptiles and mammals. Studies of microstructure of the test and tooth apparatus of fossilized animals conducted by A. P. Bystrov, remain unsurpassed in thoroughness and accuracy. He was a scientist of wide background, a consistent follower of Darwin, and an outstanding pedagogue. His book Past,

Present and Future of Man (1958) is of considerable interest. He was awarded the order of the Red Flag and medals.

Lev Borisovich Rukhin died on September 8, 1959 as a result of an accident; he was doctor of geologic-mineralogic science, professor of Leningrad State University, member of the Communist Party since 1941, an outstanding Soviet lithologist. His obituary was printed in "Izvestiya Akad. Nauk SSSR, Geologic Series", No. 3, 1960.

Konstantin Mikhaylovich Koshits died on September 20, 1959; he was the senior lecturer of the chair of petrography of the geology faculty of Leningrad State University, and member of the Communist Party of the Soviet Union since 1942. He was born on December 23(10), 1903, graduated from the geology department of Leningrad State University in 1940 and was appointed assistant to the chair of petrography in 1943. His studies dealt mainly with the problems of petrography of various regions of Kola peninsula and also with alkalic rocks and minerals of that area. In 1948 he was awarded Stalin's Prize of second degree for discovery of important iron ore deposits. K. M. Koshits was also awarded two war orders of second degree and medals.

On October 11, 1959 Boris Vasil'yevich Ivanov died; he was senior scientific worker of the Institute of Geology of Ore Deposits, Petrography, Mineralogy and Geochemistry of the Acad. of Sciences of the U. S. S. R., candidate of geologic-mineralogic sciences, and member of the Communist Party of the Soviet Union since 1942. His works on petrography of industrial stone have been widely read; see his obituary in "Izvestiya, Akad. Nauk, Geologic Series" No. 2, 1960.

On December 6, 1959 Tat'yana Konstantinovna Kozhina died; she was candidate of geologic-mineralogic sciences, junior scientific worker of the Institute of Geology of Ore Deposits, Petrography, Mineralogy and Geochemistry of the Academy of Sciences of the U. S. S. R. She was born on November 20, 1912, graduated from the Moscow Geological Survey Institute in 1936 and started working for the Institute of Geological Sciences of the Academy of Sciences of the U. S. S. R. Her studies dealt with geology of ancient formations of the Arctic and Subarctic Urals, mainly petrography of granitoids common there and on metallic ore formation.

For the work Geologic Structure and Minerals of the Watershed Part and the Eastern Slope of the Subarctic Urals T. T. Koshivina and coauthors were awarded the Kirov Prize in 1950 by the Presidium of the Academy of Sciences of the U. S. S. R.

On March 25, 1960 Iya Aleksandrovna Rukavishnikova died; she was candidate of geologic-mineralogic sciences, senior scientific worker of the Institute of Geology of Ore Deposits, Mineralogy, Petrography and Geochemistry of the Academy of Sciences of the U. S. S. R. She was born on November 3, 1905, graduated from the geological survey faculty of the Sverdlovsk Mining Institute in 1937 and began working in the Institute of Geological Sciences of the Academy of Sciences of the U. S. S. R. During the first years she studied wolframite occurrencies in granite massifs of the Urals; later her investigations dealt with the problems of ancient weathered crusts and associated minerals. I. A. Rukavishnikova was an outstanding specialist in the field of precise mineralogic analysis by optical methods. Very important was her work on oxidization zones and on geochemical mapping of polymetallic deposits of central Kazakhstan. She was awarded order "Znak Pocheta" (Sign of Esteem) and two medals.

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